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STRUCTURAL PETROLOGY OF THE BARABOO QUARTZITE . . . N. ALLEN RILEY 453

STRUCTURAL PETROLOGY OF THE PINE CREEK AREA, DICKINSON COUNTY, MICHIGAN . . . . . JAMES W. HIGGINS 476

IGNEOUS METAMORPHISM IN THE OROFINO REGION, IDAHO CLAYTON H. JOHNSON 490

THE COMPOSITION OF METEORITIC MATTER: II. THE COMPOSITION OF IRON METEORITES AND OF THE METAL PHASE OF STONY METEORITES HARRISON BROWN AND CLAIRE PATTERSON 508

UPRIGHT TRUNKS OF *NEOCALAMITES* FROM THE UPPER TRIASSIC OF WESTERN COLORADO . . . . . EDWARD L. HOLT 511

THE RATE OF ROUNDING OF BEACH BOULDERS . . . . . J. A. BARTRUM 514

THE ORIGIN AND IMPORTANCE OF THE RALEIGH GRAPHITE JOHN W. HARRINGTON 516

DIABASE DIKES OF THE FRANKLIN FURNACE, NEW JERSEY, QUADRANGLE CHARLES MILTON 522

REVIEWS

Dating the Past: An Introduction to Geochronology, FREDERICK E. ZEUNER (Ernst Antevs) . . . 527

Glacial Geology and the Pleistocene Epoch, RICHARD FOSTER FLINT (Leland Horberg) 530

Geological Expedition to the Lesser Sunda Islands under the Leadership of H. A. Brouwer (R.T.C.) . 531

PUBLICATIONS RECEIVED 532

INDEX 535

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STRUCTURAL PETROLOGY OF THE BARABOO QUARTZITE

N. ALLEN RILEY
Department of Geology, University of Chicago

ABSTRACT

The microstructures of the Baraboo quartzite of Columbia and Sauk counties, Wisconsin, have been studied from oriented hand specimens, polished sections, and thin sections. Fabric axes, determined from the pole patterns of the deformation lamellae, vary consistently over the region and are not related to the bedding or the macrofoliation. Because of this variation, the deformation lamellae have been assigned to a deformation later than the major folding of the region. In addition, the character of the deformation lamellae themselves and the type of pole pattern have been used as a qualitative index of the intensity of the deformation. Microfractures are common features of the quartzite; they exhibit a strong preferred orientation and are geometrically related to the deformation lamellae. Quartz deformation bands are described, named, and correlated with bedding-plane shear. Quartz optic-axes fabrics are weak and difficult to interpret.

INTRODUCTION

The present work is a field study of the outcrops and laboratory examination of oriented hand samples, polished sections, and thin sections of the Baraboo quartzite of Columbia and Sauk counties, Wisconsin. It is not my purpose to attempt a solution of the fundamental problem of the dynamics of quartz deformation or the mechanics of tectonic movements but rather to shed further light on the qualitative description of the processes of deformation and their correlation with the macrostructures observed in the field.

GENERAL GEOLOGIC DESCRIPTION OF THE BARABOO REGION

The Baraboo quartzite crops out in Columbia and Sauk counties, Wisconsin. Its structure (Fig. 1) is a complex canoe-shaped synclinorium, whose major axis strikes approximately N. 70° E. and is roughly 25 miles east-west and 9 miles north-south. The outcrop area can be conveniently divided into a north, a south, and a west range. In the north range the bedding is mainly vertical, striking east-west, with the tops of the beds to the south; in the south range it dips 20°–40° to the north, striking N. 70° E.; the much shorter west range is an area of complex, rather right, east-west folds, plunging eastward. In the western half of the south range, a series of gentle folds, trending N. 60° E., truncates the larger structure.

The Baraboo area is of considerable geologic interest because of excellent exposures of axial-plane cleavage, drag folds, etc. Numerous papers concerning the various aspects of the region have appeared in the literature. The most comprehensive treatise on the area and one with considerable emphasis on the structure of the quartzite ranges ap-

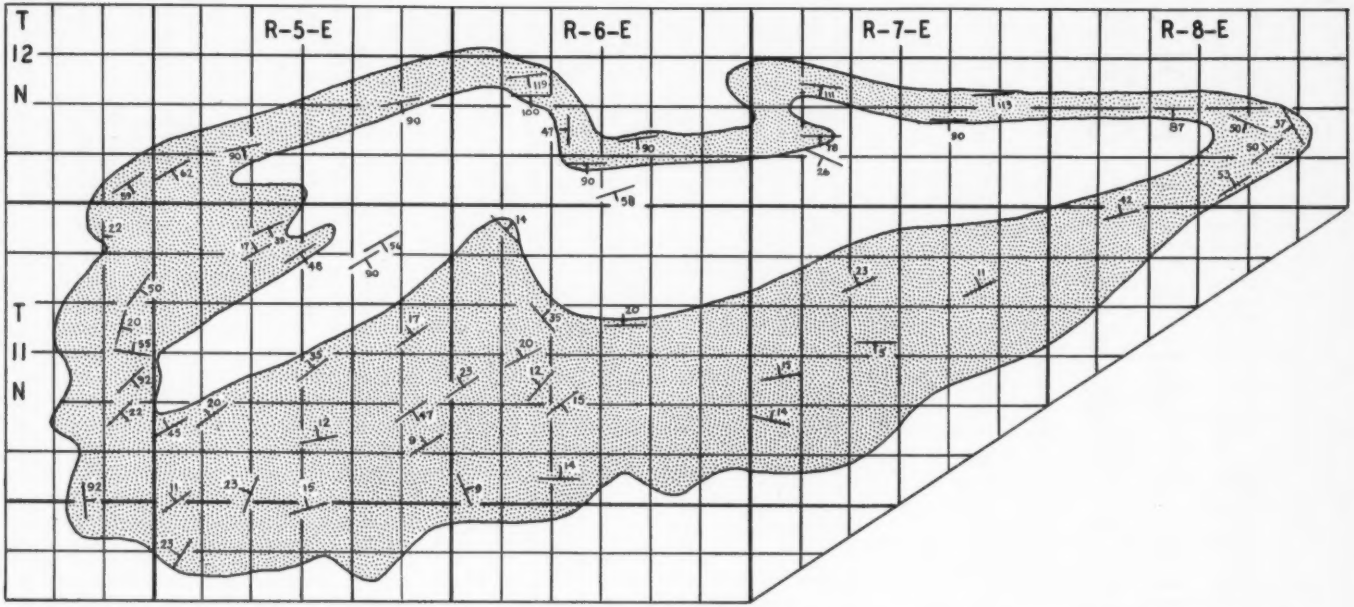


FIG. 1.—Generalized outcrop area of the Baraboo quartzite (modified from A. Leith), showing bedding orientation for selected representative locations. The four outcrops in the interior of the basin are possible outcrops of the Dake quartzite.

peared in 1904 by Samuel Weidman.[1] For the details of the region reference must be made to Weidman's work. It is sufficient to say that Weidman correctly placed the Baraboo quartzite in the Huronian, described the structure as a complex synclinorium, and divided the Huronian sediments into three formations: the Baraboo quartzite, the Seeley slate, and the Freedom iron formation. The latter two formations, which are above the Baraboo quartzite, do not crop out in the region and were described by Weidman from mine workings.

Andrew Leith[2] restudied the earlier unpublished commercial reports of C. L. Dake and in 1935, on the basis of Dake's descriptions of drill cores, divided the Baraboo Huronian sediments into an upper and lower part. Leith's pre-Cambrian column is as follows:

Algonkian	Upper Baraboo series	Rowley Creek slate
		Dake quartzite
	Unconformity (?)	
	Lower Baraboo series	Freedom formation
		Seeley slate
		Baraboo quartzite
	Unconformity	
Archean		Rhyolite, tuff, granite (some granite may be intrusive into the Baraboo series)

A short description of the formations of the lower Baraboo series, taken from Weidman's[3] publication, is given below:

The Baraboo quartzite exceeds 4,000 feet in thickness and is composed almost entirely of monocrystalline sedimentary quartz grains with varying admixtures of fine-grained sericitic material. Interstitial iron oxide imparts a reddish color to much of the formation. Detrital hematite and zircon grains are locally important. All large quarries in the Baraboo quartzite exhibit thin layers of almost pure sericitic material, varying in thickness from a few centimeters to several feet, interbedded with massive quartzite. Some of these thin beds are pinched into lens-shaped bodies. A poorly developed basal conglomerate is present in one known locality. Intraformational pebble bands of vein quartz are common features of the massive quartzite.

The Seeley slate is not known to crop out in the region and is recognized only in mine workings and drill holes. Its material is observable at present on the dumps of old mines and is a finely banded, soft slate, exhibiting excellent axial-plane cleavage. The thickness of the formation is unknown.

The Freedom formation is the iron-bearing formation of the region. It does not crop out in the area. Weidman describes the formation as composed of slate, chert, dolomite, iron ore, and all

[1] "The Baraboo Iron Bearing District," *Wis. Geol. Nat. Hist. Surv., Bull. XIII, Econ. Ser. No. 8* (1904), pp. 1–190.

[2] "The pre-Cambrian of the Lake Superior Region, the Baraboo District, and Other Isolated Areas in the Upper Mississippi Valley," *Ninth Ann. Field Conf. Kan. Geol. Soc.* (Wichita, Kan.: Kansas Geological Society, 1935), pp. 320–32.

[3] Pp. 22–29 of ftn. 1.

gradations among these types. The thickness of the formation is unknown but is estimated by Weidman to exceed 1,000 feet.

Leith,[4] utilizing Dake's data, named and described the formations of the Upper Baraboo series as follows:

> Rowley Creek slate: This slate does not outcrop. . . . . It occupies the middle of the synclinal trough. . . . . Its thickness varies between 40 and 149 feet. . . . .
>
> Dake Quartzite: This formation was encountered at 42 diamond drill holes. . . . . It varies in thickness up to a maximum of 214 feet. . . . . Lithologically, it consists of a coarse grained quartzite which contains a large amount of sericite and chlorite as a matrix to the quartz grains. A large portion of the formation is coarsely conglomeratic with large, and sometimes angular, pebbles. . . . .
>
> The quartzite is not known to outcrop, but the writer believes that one, and possibly several, surface exposures formerly thought to be Baraboo quartzite are of Dake quartzite. . . . .

Leith points out that two features of the structure indicate an unconformity between the Upper and the Lower Baraboo series. The drill records show that the Dake quartzite bevels the underlying Freedom formation; and, if the surface outcrops of the Dake are correctly identified, they dip at gentler angles than does the underlying exposed Baraboo quartzite. An example of the latter is Sample 55, Figure 14, which is the "type" surface exposure of the Dake. The bedding here dips approximately 25° southward, while a short distance to the north the bedding in the Baraboo quartzite is roughly vertical (the symbols used in all illustrations are explained in Fig. 2).

An attempt was made in the present study to determine fabric differences between the Dake and the Baraboo quartzites in the above outcrops in the hope that fabric differences would show the unconformity. No evidence was obtained from the mircofabric studied either to substantiate or to refute the presence of the unconformity. The fabric axes determined from the pattern of the poles of the deformation lamellae (discussed in detail later in the paper) of both outcrops are similar.

Another problem associated with the structure is the determination of the number of deformations that the Lower Baraboo series has undergone. To begin with, it appears unlikely that the complexities of structure encountered in the Baraboo quartzite could be derived in one simple deformation. Evidence is submitted that the Lower Baraboo series has suffered at least two deformations. If the angular unconformity between the Upper and the Lower series is admitted, then the region has been subjected to at least three deformations since Lower Baraboo times.

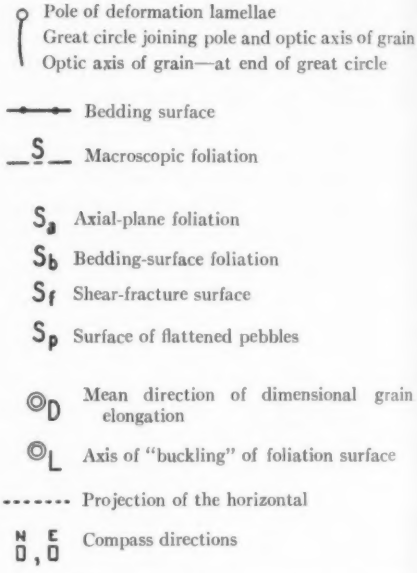


FIG. 2.—Symbols used on all maps and projections

[4] Pp. 329–30 of ftn. 2.

Volcanics, basal to the Baraboo series, occur locally around the perimeter of the synclinorium. These outcrops were not studied in detail and will not be discussed in this paper.

Resting unconformably on the Baraboo series are Upper Cambrian sands. They are roughly horizontal and crop out both in the interior of the synclinorium formed by the Baraboo series and on all sides. Several small-scale faults are found in these Cambrian sands in the interior of the synclinorium. The displacement along the fault planes can be measured in inches, the strike is roughly east-west, and the downthrow side is toward the interior of the basin. For these reasons the Paleozoic faulting is attributed to differential compaction of the sediments, and no major Paleozoic or later deformation is recognized.

The difficulties in determining the field structural relations in the Baraboo quartzite are the absence of any traceable horizon within the quartzite, the difficulty in recognizing tops and bottoms of bedding surfaces, and, in many examples, difficulties even in determining bedding surfaces. The quartzite varies tremendously in its degree of deformation over small distances. Many massive outcrops exhibit apparently undeformed ripple-marked surfaces, while a few tens of feet away stratigraphically, in more sericitic layers, large-scale drag folding and pronounced boudinage occur.

An enumeration of the principal macrostructural features follows:

1. Bedding surfaces are identified both where the change in lithology with respect to amounts of admixed sericite is gradational and where it is abrupt. Cross-bedding surfaces are easily confused with true bedding in small outcrops, and small errors in the orientation of the bedding surfaces are expected.

2. Axial-plane foliation is pronounced in the thicker layers, which contain considerable amounts of sericite. It is found only in the north range, where the bedding is close to vertical. The orientation of this foliation is roughly N. 60°–80° E., dipping 40°–60° N.

3. Bedding-plane foliation is found in the more massive layers in the north range and is the principal foliation of the south range.

4. Shear-fracture surfaces can be divided into the following types:

a) Shear fractures in massive quartzite that are observed to pass into axial-plane foliation in the more sericitic layers.

b) Chevron folding of the axial-plane foliation is locally observed. This type of structure has been described by Barrow,[5] who calls these acute plications "buckling." Fracture at the crests and troughs of the buckling and small displacements along the fracture surfaces are readily observable. The amplitude and wave length of the buckling observed varies between 0.05 and 2.5 cm.

c) Shear fractures, discontinuous and continuous, in the more massive members transcend all other structures in the south and west ranges. These are roughly vertical and trend east-west. The formation of these shear fractures was apparently subsequent to all major deformation.

5. Metamorphic surfaces consisting of flattened pebbles are prominent in the outcrop at the locality of Sample 55, Figure 14. This is a quartzite outcrop which Leith said might be an exposure of the Dake formation.

6. Macrolineation is locally evidenced by the following factors:

[5] "On Buckled Folding," *Rept. Brit. Assoc. Adv. Sci.*, Vol. LXXXII (1912), pp. 473–74.

a) Intersection of axial-plane foliation and bedding
b) Intersection of shear fractures and bedding
c) Crests and troughs of the buckling
d) Slickensides

7. Quartz veining of several forms is recognized:

a) In the north and west ranges, veins of quartz varying in thickness from a few millimeters to about 60 cm. are common. These veins follow joint patterns and transect all metamorphic structures.

b) In highly deformed quartzites, deformed quartz veins are present. Some of these deformed veins are cut by planar veins.

c) Ladder veins of quartz are found in the competent members of large-scale drag folds. They can be interpreted as vein filling of tension fractures which are roughly perpendicular to the bedding.

d) S-shaped gash veins of quartz are very common. They have not been interpreted and have not been related to the regional structure

8. Large-scale drag folds are found in three outcrops near the top of the exposed Baraboo formation in the south range, near where Sample 20, Figure 14, was collected. Boudinage is well marked, both parallel and perpendicular to the dip of the bedding.

9. No large-scale faults are recognized in the field. Leith[6] places a north-south fault between Samples 58 and 38 of Figure 14. This fault was apparently based on drill records. If a large-scale fault is present, it is not represented in the pattern of the deformation lamellae, because the fabric axes determined from the pattern of these lamellae are similarly oriented in both the above samples, as well as in those samples farther west from the supposed fault. (Fig. 14.)

[6] Pp. 329 of ftn. 2.

Consistently throughout the area, roughly vertical, north-south shear joints or perhaps shear fractures are encountered. Slickensides on these surfaces are rare but are found frequently enough to assume some north-south adjustments by fracturing. It is interesting to note that these surfaces are subparallel to the *a*–*b*-fabric planes determined from the deformation lamellae.

MICROSTRUCTURES

I. SINGLE GRAINS

The principal intragranular microstructures include the following:

1. Secondary quartz enlargement of the original quartz grains is visible in many thin sections. The original boundary is visible because a thin film of impurities at the original grain surface is incased by the secondary enlargement. The above is true only in samples containing small amounts of sericite. Samples containing large amounts of sericite, however, rarely exhibit original boundaries of the quartz grains, but each grain is completely surrounded by sericite grains. These sericite grains appear to penetrate the boundaries of the quartz grains as if sericite had replaced quartz. An original boundary is rarely observed, but when it is, it can be seen that the original quartz grain has grown by secondary quartz enlargement and has incased the adjacent sericite grains.

2. Deformation, or Böhm, lamellae are frequently observed structures within individual grains. They commonly occur in the central portions of the grains and appear as a set of closely spaced subparallel planes which are slightly lens-shaped in cross section. Infrequently, two sets of deformation lamellae are found in the same grain. A historical summary of previous work on deformation lamellae and

a careful description of them has been given by Earl Ingerson and O. F. Tuttle.[7]

3. Undulatory extinction is present in nearly all quartz grains observed. The magnitude of this change in optic orientation varies tremendously from grain to grain in one sample and in different samples. In addition to this continuously varying position of the optic axis in a particular grain, discontinuities occur which divide the grain into bands subparallel to the optic axis. Anna Hietanen[8] has extensively described deformed quartz grains from the Finnish quartzites, and she considers the undulatory extinction as being due to bend-gliding along the basal pinacoid and the discontinuities subparallel to the optic axis as actual rupture planes.

4. Undulatory extinction discontinuities with apparently random orientation are observed as thin bands across the grain. They have a slightly different optical orientation and/or index of refraction from the rest of the grain and are, therefore, seen only when the grain is very close to an extinction position. Two sets of bands, whose traces tend to be perpendicular, are present in most grains showing the banding. These bands agree in appearance with structures in vein quartz observed by Sidney Adams[9] and are termed "deformation bands," following the usage in metallurgy.[10]

5. Sutured boundaries between quartz grains are common. An examination of variously deformed quartzites shows a wide variation in the character of the sutured boundaries. The individual sutures are definitely different in different samples; they are not necessarily perpendicular to the interface between grains; they vary in an individual grain according to the optic orientation of the neighboring grain, which makes the interface; and they exhibit marked differences in the amount of interpenetration. In examples in which the interpenetration is large, the sutured parts of both grains differ considerably in optical orientation from their parent-grains. All variations between no change in optic orientation of the sutured parts of the grains to considerable change are observed. This suturing of grain boundaries and then rotation of the sutured parts with recrystallization could be major processes in the breakdown of large grains. In addition to the above, differences in the amount of interpenetration of the sutures are observed in variously oriented sections of the same sample. This observation indicates that the character of the sutures is probably related to the major stress pattern. No investigation of suturing other than the above general observations has been made in this study.

Much less frequent than suturing is the "pressing-in" of one side of a particular grain into an adjacent one. In some examples it is possible to observe the complete original boundary of one of the grains and the partial original boundary of the other grain.

6. Breakdown of the original grains by deformation can be observed as having taken place

a) Along planes subparallel to the optic axis, where the original grains break down into needle-like fragments exhibiting strong undulatory extinction.

[7] "Relations of Lamellae and Crystallography of Quartz and Fabric Directions in Some Deformed Rocks," *Amer. Geophys. Union Trans.*, Vol. XXVI (1945), pp. 95–105.

[8] "On the Petrology of the Finnish Quartzites," *Bull. Comm. géol. Finlande, No. 122* (1938), pp. 1–118.

[9] "A Microscopic Study of Vein Quartz," *Econ. Geol.*, Vol. XV (1920), p. 648.

[10] Charles S. Barrett, *Structure of Metals* (New York: McGraw-Hill Book Co., 1943), pp. 305–7.

b) Along the sets of deformation bands, where the whole grain breaks down into small angular units that are roughly equant in cross section and vary slightly from one another in optic orientation. The end-result of this breakdown could result in the texture called "crushed quartz" by R. E. Fellows.[11]

c) At the boundaries among grains, where small sutured fragments are detached from their parent-grains.

d) In the interior and at the boundaries among grains, where recrystallization like that illustrated by J. A. Hedvall[12] occurs, that is, growth of new crystals at the expense of the old ones.

7. Small subhedral grains are found in the interstices among the larger grains. They show no trace of a rounded surface, have very irregular but equant shapes, and appear as if they had grown around new nuclei, that is, had recrystallized.

8. In some of the quartzites that contain negligible amounts of sericite and have suffered extensive deformation, as shown by the partial breakdown of the larger grains, very elongate quartz grains are observed. Rarely, one of these has an original sedimentary boundary, indicating that it formed from an original grain and was not the result of growth from a new nucleus or of coalescence of several grains. An elongate grain commonly exhibits marked discontinuities in the undulatory extinction subparallel to the optic axis when it is subparallel to the dimensional elongation. On the other hand, if the grain is dimensionally elongated subperpendicular to the optic axis, it may show well-developed deformation lamellae in addition to the discontinuities in the undulatory extinction. Samples of quartzites containing large amounts of sericite display little elongation of the original quartz grains of the sand size. These grains have reacted to the stress like a rigid body in a fluid medium.

[11] "Recrystallization and Flowage in Appalachian Quartzite," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), p. 1414.

[12] *Reaktionsfähigkeit fester Stoffe* (Leipzig: Verlag von Johann Ambrosius Barth, 1938), p. 53.

II. POLYCRYSTALLINE AGGREGATES

The principal intergranular microstructures of the Baraboo quartzite include the following:

1. Planar structures passing through more than one grain without deviation at the grain boundaries are commonly observed. Some of these are marked by thin veinlets of very fine-grained sericite. In most examples, however, they are discontinuous planes of inclusions. In all examples, continuous or discontinuous, these planes show a pronounced parallelism. More than one set of such structures can be identified in some samples. These planar structures are termed "microfractures."

2. Quartz microveins are common. They occur as planar structures and might be thought of as having grown along the microfracture surfaces described above. In addition, quartz microveins are found that could be described as "contorted." Both types have been found in the same thin section, with the planar veins cutting the contorted veins.

3. Microfoliation surfaces are visible in samples containing large amounts of sericite as concentrates of sericite cutting across relic bedding but not cutting through original quartz-grain boundaries.

4. Original sedimentary structures commonly observed in thin sections are the following:

a) Laminae of heavy mineral concentrates are recognizable in a large number of samples. These laminae are valuable

field aids for the determination of the sedimentation surface. They rarely exceed a few millimeters in thickness and are composed exclusively of hematite and zircon grains. Sufficient zircon grains could be found for a regional study of the zircon fabric.

b) Graded bedding is prominent on a microscale, and in most examples one graded unit represents one cross-bedding lamina.

c) Ripples and truncated ripples are visible in thin section by slight changes in grain size or by variation in abundance of hematite, zircon, or sericite.

d) Some polycrystalline (sedimentary) quartz grains are present in the more conglomeratic phases of the Baraboo quartzite. Some of the individual grains show a much higher degree of metamorphism as well as a greater variation in degree of metamorphism from grain to grain than is true for monocrystalline grains of the same size in the same sample. This variation indicates a source region or regions in which the quartzites differed widely in degree of metamorphism. As will be shown later in this paper, the region may be as small as the outcrop area of the Baraboo quartzite and still show this wide range in degree of metamorphism.

DETAILS OF THE MICROSTRUCTURE

I. DEFORMATION LAMELLAE

The deformation lamellae observed in the Baraboo quartzite agree in the main with those described by Ingerson and Tuttle.[13] Contrary to their findings in the Ajibik quartzite and in a biotite gneiss from near Washington, D.C., however, deformation lamellae are common in the Baraboo quartzite and are relatively easy to observe in the majority of examples. In samples containing so much sericitic material that the individual quartz grains are not in contact with one another, however, the readjustments during deformation have been made by the sericite grains leaving the quartz apparently undeformed and without deformation lamellae.

Various interpretations have been given to the deformation lamellae, but, following Ingerson and Tuttle, they can all be classified as: (1) planes of liquid inclusions, (2) healed fractures, and (3) translation glide-planes. Their exact nature is not known; their relation to quartz structure is obscure; and their orientation with respect to the major-stress axes is based on field evidence. Deformation lamellae have not been observed in laboratory deformation of quartz.

H. W. Fairbairn[14] measured the angle between the poles of the deformation lamellae and the optic axes of the same grain and found a total variation of from 3° to 85° with maxima at 18° and 27°. He interprets the deformation lamellae as translation glide-planes containing the glide-line $[m:r]$.

Ingerson and Tuttle[15] found the angle between the optic axis and the pole of the deformation lamellae to vary with the angular distance from the optic axis of the grain to the *B*-fabric axis ("*B*" is written "*b*" in this paper). In addition, they tabulated the solid angle formed by the planes containing the optic axis and poles to two sets of deformation lamellae from the same grain. If deformation lamellae are translation glide-planes containing the glide-direction $[m:r]$, then this angle is 60°. No evidence of this 60° angle was found.

[13] Pp. 95–96 of ftn. 7.

[14] "Quartz Deformation and Crystal Structure," *Amer. Min.*, Vol. XXIV (1939), pp. 351–68.

[15] P. 101 of ftn. 7.

The deformation lamellae do form more frequently at low angles to the basal pinacoid. The distribution of angles (optic axis to pole of deformation lamellae) of Ingerson and Tuttle and of Fairbairn, tabulated in Figures 10 and 11 of Ingerson and Tuttle,[16] were substantiated by measurements on the Baraboo quartzite. In addition, the perpendicular distance between individual lamellae was measured. The spacing varied between 2.5×10^{-4} and 17×10^{-4} cm., with a mean value of 8.3×10^{-4} cm. These values agree in magnitude with the observed spacing of slip-planes in metallic crystals.

Ingerson and Tuttle relate the deformation lamellae to the fabric axes; that is, the intermediate-strain axis is parallel to the lamellae (or the lamellae poles lie in the *a*–*c*-fabric plane). Ingerson and Tuttle have further pointed out that if the lamella pole and the optic axis of quartz are plotted for an individual grain and are joined by a great circle, the pole to the lamella always lies between the optic axis and the *a*-fabric axis (Fig. 3), that is, the great circles joining optic axes and deformation-lamellae poles always "point" to the *a*-fabric axis.

[16] P. 99 of ftn. 7.

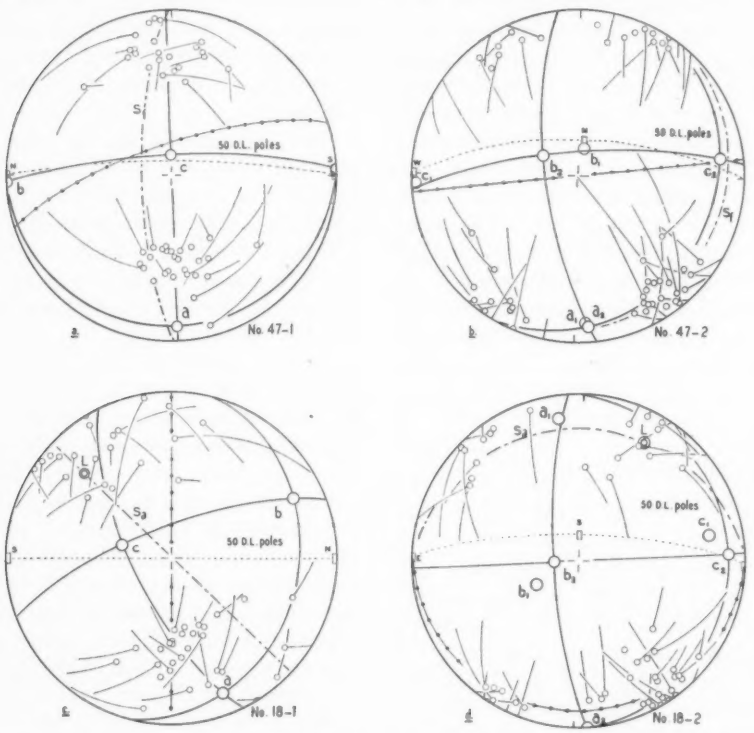


FIG. 3.—From Sample 47 two perpendicular sections have been cut, 47-1 and 47-2. The *a*-, *b*-, and *c*-fabric axes were determined independently in each section from the pattern of the poles of the deformation lamellae. The fabric axes determined from 47-1 are plotted on the projection for 47-2 and labeled as a_1, b_1, c_1. The above also applies to the two sections from Sample 18. The top of the bedding is to the south in Sample 18.

No consistent relationship was found in the Baraboo quartzite between the *a*–*b*-fabric plane determined in the field from the macrofoliation surfaces and the fabric axes determined from the pattern of the deformation lamellae. To avoid confusion of terminology, the *a*-, *b*-, and *c*-fabric axes are directions determined according to Ingerson and Tuttle from the pattern of the deformation lamellae. Megascopic foliation surfaces in all examples are here designed as *S*-surfaces, with subscripts to denote type of *S*-surface. Figure 14 shows the variation in the fabric axes as determined from the deformation lamellae over the outcrop area of the Baraboo quartzite. The interpretation of this pattern will be discussed later.

In most samples the poles of the deformation lamellae are concentrated in two small maxima, and the *a*-, *b*-, and *c*-fabric axes can be critically located with reproducible results from variously oriented thin sections of the same sample and from different areas of the same thin sections. Two examples of this for a pair of mutually perpendicular thin sections from the same sample are given in Figure 3.

As deformation of the aggregate increases, an individual grain will be subjected to a changing-stress field. Deformation of a particular grain and deformation of its neighboring grains will alter the local stress pattern. A wide range in character, strength, and deformation of the deformation lamellae themselves is observed in an increasing-deformation field. Increasing deformation, of course, applies to a sequence of changes as observed in a series of rock specimens. In addition to the location of the *a*-, *b*-, and *c*-fabric axes, therefore, considerable evidence concerning the deformation can be obtained from the variation and character of the deformation lamellae. Ingerson and Tuttle[17] indicated the above but did not elaborate.

The following detailed observations serve to illustrate the variations in the character of the deformation lamellae and in the pole pattern of the deformation lamellae. They are from samples containing negligible amounts of sericitic material, unless otherwise indicated.

1. The samples which exhibit the most precisely oriented pattern of poles of the deformation lamellae are those whose quartz grains show slight undulatory extinction, no noticeable dimensional elongation, distinct original boundaries with secondary enlargement, and rare other intragranular discontinuities. From the above, it appears that in an increasing-deformation field these lamellae are the first deformation structures to appear. Many examples can be found in which the deformation lamellae cross the boundary from the original grain into the region of secondary enlargement. Most of the deformation lamellae, therefore, followed the secondary enlargement. The exact relations between the development of undulatory extinction and the formation of deformation lamellae is difficult to evaluate, but grains exhibiting deformation lamellae also show some slight undulatory extinction. Sample 47, whose pole pattern of deformation lamellae is shown in Figure 3, is an excellent example of the above.

2. Curved deformation lamellae are found in samples exhibiting strong undulatory extinction, which has discontinuities subparallel to the optic axis, seldom observed original grain boundaries, and a noticeable grain dimensional elongation. The orientations of the great circles joining the optic axes and the poles of the deformation lamellae vary from the

[17] P. 105 of ftn. 7.

ideal examples shown in Figure 3. Figure 4, Sample 9, is the pole pattern of a sample showing the above features.

3. One large diffuse maximum of the poles of the deformation lamellae and several smaller minor maxima are shown by samples exhibiting the partial breakdown of the original grains to smaller elements, rare original grain boundaries, and a pronounced grain dimensional elongation. The *a*-, *b*-, and *c*-fabric axes are difficult to locate from the pole pattern. The deformation lamellae in the elongate grains are oriented about 30° to the elongation (Fig. 5, Sample 28). The unequal development of the pole maxima of the deformation lamellae could be explained in the following ways:

a) The deformation lamellae tend to form subperpendicular to the *c*-axis of the grain. With certain patterns of preferred orientation of the *c*-axes, one set of deformation lamellae would be more common than the other. Diagrams showing a definite *c*-axis preferred orientation can possibly be correlated with the deformation lamellae pole pattern (Fig. 5).

b) The deformation lamellae are re-

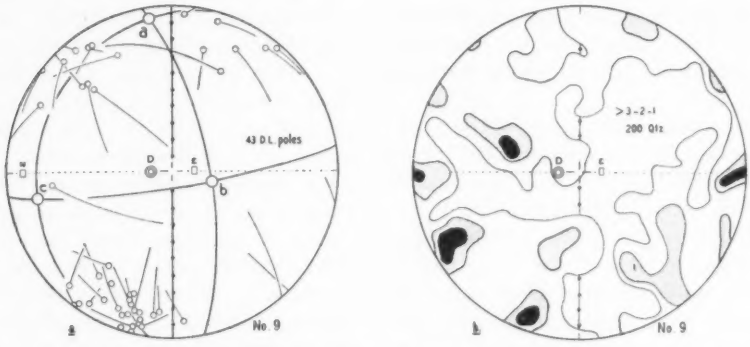


FIG. 4.—Pattern of poles of deformation lamellae in which one maximum is stronger than the other. A weak optic-axes fabric from random grains indicates a possible correlation.

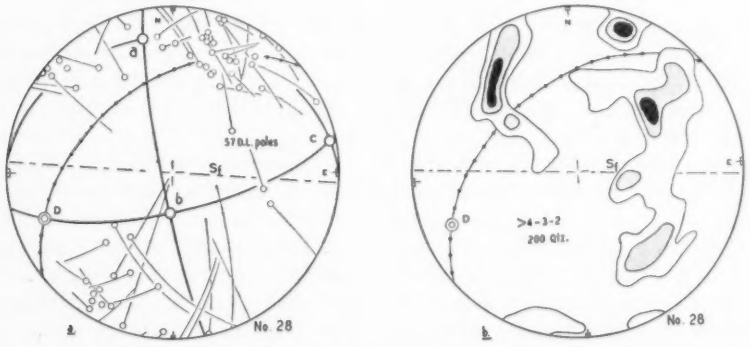


FIG. 5.—One large diffuse maximum of poles of the deformation lamellae, one or possibly two small maxima, and many aberrant lamellae make the choice of the fabric axes difficult. The pronounced dimensional elongation and the definite preferred orientation of the optic axes is not directly correlative with the deformation-lamellae pattern.

lated to the macroshear surfaces, one of which is better developed than the other. No constant relation was found between macrofoliation surfaces and the planes of deformation lamellae.

c) The deformation lamellae are related to the dimensional grain elongation, i.e., metamorphic dimensional grain elongation, which, in turn, is related to the optic-axes fabric. In rocks showing the most highly oriented and symmetrical pattern of poles of deformation lamellae, the deformation lamellae appear to form subperpendicular to the long-dimensional axis of the grains. This is "according to Hoyle" because grains which are nonequant suffer, on the average, the greatest bending parallel to their long dimension. If actual movement, either by gliding or by fracture, takes place along the deformation lamellae, then, under continued deformation, the individual grains are elongated in a direction in the plane of the deformation lamellae. With infinite elongation the greatest dimensional elongation of the grains would become parallel to the set of deformation lamellae causing the elongation. This hypothesis is crudely verified by measurements of the angle between the elongation and the deformation lamellae in specimens exhibiting various amounts of grain elongation. In addition, the grain elongation can be expressed as a flattening, which can be taken as two principal directions of elongation. The better-developed system of deformation lamellae would then logically be subparallel to the greater axis of elongation. In only one example—Sample 55, Figure 6—flattening of the quartzite grains was observed, and the above relation was not verified. In this example, however, one set of deformation lamellae is perpendicular to the flattening and one subparallel.

The logical conclusions are twofold. In very elongate grains in which the deformation lamellae are oriented at small angles to the dimensional elongation, the observed dimensional elongation could have been accomplished by gliding along the deformation lamellae. On the other hand, if the deformation lamellae are roughly subperpendicular to the long-dimensional elongation of the grains, then the deformation lamellae were formed after the event which caused the grain to be dimensionally elongated.

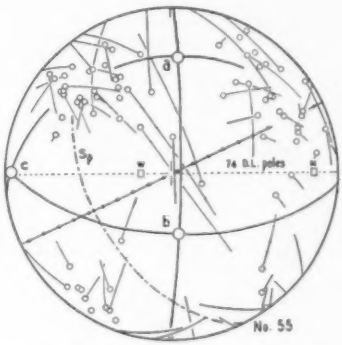


FIG. 6.—Conglomeratic quartzite (Dake?) in which the grains are flattened. In the surface of flattening, S_p, the greater elongation approximately coincides with the bedding. The deformation lamellae and the flattening are possibly related in that one pole maximum coincides with the surface of flattening and the other is approximately perpendicular.

The above interpretations are undoubtedly related but are difficult to demonstrate because neither the orientation of the forces applied nor the number of deformations is known and also because the measurement of average grain elongation in the thin section is difficult and based on the assumption that the grains are cylindrical in shape.[18] Since the major axis of metamoprhic dimensional grain elongation apparently lies

[18] Kurt E. Lowe, "A Graphical Solution for Certain Problems of Linear Structure," *Amer. Min.*, Vol. XXXI (1946), pp. 425–34.

close to the bedding surfaces (Fig. 14, Samples 9, 55, 35, 28, 3), it seems logical to assume that the stress causing the elongation has acted along the bedding surfaces. The pattern of deformation lamellae should then be related to the bedding and to the axial-plane foliation, which is, in turn, related to the bedding. No such relation is found consistently over the region. A possible explanation is that the pattern of deformation lamellae represents a later deformation imprinted

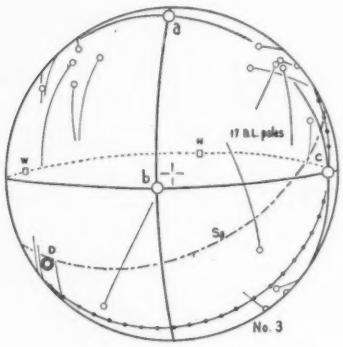


FIG. 7.—Pole pattern of deformation lamellae from a sericite schist layer. The few deformation lamellae observable adequately bring out the pattern. No relation is apparent between the pole pattern and the axial-plane foliation, S_a, observed from the outcrop, in the hand sample, and in the thin section. For another example of the same rock type, see Sample 18, Fig. 3*c* and *d*.

on an earlier pattern of metamorphic structures. Samples 3 and 18, Figures 7 and 3, are from sericite schist members of the Baraboo quartzite which exhibit marked axial-plane foliation. The fabric planes determined from the deformation lamellae do not bear any detectable relation to this foliation. In their examples Ingerson and Tuttle have demonstrated that the fabric axes determined from the deformation lamellae are related to the macrofabric axes, that is, the *a*–*b*-fabric plane determined from the deformation lamellae is parallel to the axial-plane foliation. If we accept their data, then the fabric axes determined from the deformation lamellae in the Baraboo quartzite are related to a different system of forces than those which resulted in the axial-plane foliation.

4. In several pole patterns of deformation lamellae (Samples 5 and 6 in Fig. 14; the pole pattern for Sample 6 is shown in Fig. 8), the poles form a girdle about the *a*-fabric axis. This girdle has an apical angle of about 80°. The pattern, therefore, has one degree of freedom about the *a*-fabric axis; thus the *b*- and *c*-fabric axes cannot be uniquely determined. Either of two explanations appears possible: that this pattern represents true biaxial strain or that this pattern represents two or more deformations, during which the *a*-fabric has remained essentially constant in direction but with various directions of the *b*- and *c*-fabric axes.

5. Deformation lamellae whose planes are marked by a thin film of brown material are recognized. Ingerson and Tuttle,[19] when they can detect no difference in index of refraction between the lamellae and the rest of the grain, call this type of lamella "relict." Very minute changes in the density of the brown film in one lamella can be traced through the set of the lamellae, subperpendicular to the lamellae, subparallel to the optic axis of the grain, and parallel to the discontinuities in the undulatory extinction. Hietanen describes a quartzite in which the deformation lamellae, termed "Böhm lamellae" in her paper, are curved, showing a wavelike pattern. Her diagrams[20] illustrate the change in texture of the brownish material along the lamellae, as described above, at the positions of curvature of the lamellae. Curvature of the deformation lamellae in

[19] P. 95 of ftn. 7. [20] Pp. 39–40 of ftn. 8.

the Baraboo quartzite and/or changes in the density of the brown film along the lamellae coincide with discontinuities in the undulatory extinction subparallel to the optic axis.

Hedvall and Sjöman[21] have shown that, at temperatures above the inversion point of quartz, Fe_2O_3 diffuses into the quartz lattice at a rate measurable in the laboratory. With sufficient time and perhaps in the presence of impurities, it is entirely possible that appreciable diffusion occurs at lower temperatures and that the brown films marking the deformation lamellae have originated in that manner.

6. Two sets of deformation lamellae rarely occur in a single grain. When observed, one set is more pronounced and appears to cut the other set. This observation does not necessarily imply two deformations, because two sets of deformation lamellae in the same grain could be explained as the result of local perturbations in the stress field during the same deformation.

If the relict deformation lamellae could be separated in a particular sample from deformation lamellae not showing the brown film and if the relict lamellae were cut by the nonrelict lamellae and if the pole patterns of the two types of lamellae were consistently different and if the regional configuration of the rocks could be correlated with two distinct deformation-lamellae patterns, then perhaps multiple deformations could be inferred. A lengthy attempt was made to achieve this result, but without success.

Summarizing the foregoing, it is possible to achieve a qualitative estimate of the intensity of the local deformation from the pattern of the deformation-lamellae poles and from the character of the deformation lamellae themselves. In an increasing intensity of deformation the following varieties appear:

1. *Straight deformation lamellae.*—These are usually in the central portions of the grains; and they are roughly perpendicular to the long axes of the grains; the pole pattern of the deformation lamellae is clear cut, with two well-developed maxima approximately 80° apart; the pole of a particular set of deformation lamellae always lies between the *a*-fabric axis and the optic axis of the particular grain.

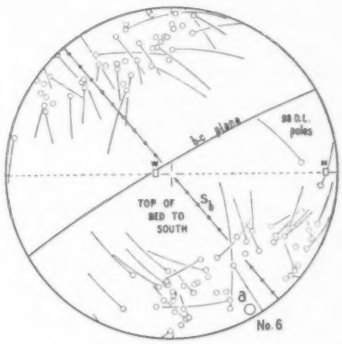


FIG. 8.—Pattern of poles of deformation lamellae in which the poles are distributed in a small circle about the *a*-fabric axis.

2. *Curved deformation lamellae.*—These lamellae appear with the curvature along planes subparallel to the optic axis and parallel to the discontinuities in undulatory extinction; the relation between the deformation lamellae and dimensional grain elongation is obscure; the pole pattern of the deformation lamellae is more diffuse than in the above examples, but the two maxima are clearly recognizable.

3. *Curved deformation lamellae as above.*—Dimensional grain elongations tend to be at small angles to the plane of the deformation lamellae; the pattern of poles to the deformation lamellae is dif-

[21] P. 212 of ftn. 12.

fuse, so that the fabric axes are determined with considerable question.

The above analysis applies only to "pure" quartzites, that is, each quartz grain is bounded by quartz grains with only small amounts of other material, such as sericite, in the interstices among the quartz grains. The stresses are then transmitted principally from quartz grain to quartz grain. As the amount of sericite, for example, is increased, part of the change in shape of the deforming aggregate is borne by the sericite. With sufficient admixture of sericite, such that the quartz grains are no longer in contact with one another, the major portion of the change in shape of the aggregate in deformation is accomplished by change in the sericite, the quartz grains acting as rigid bodies in a viscous fluid.

The study of the deformation lamellae in the Baraboo quartzite indicates the need for a better understanding of the conditions of their formation and of their exact nature. Only by producing them experimentally can the relation between the forces and the ensuing pattern be determined. If this could be accomplished, deformation lamellae should be one of the more powerful tools in the study of deformed rocks.

II. DEFORMATION BANDS

A detailed study of the variation in the orientation of the optic axis in individual quartz grains of the Finnish quartzites has been made by Hietanen.[22] She plotted the orientation of the optic axis of a particular grain on an enlarged trace of the grain outline. These plots show the angular limits of the undulatory extinction and demonstrate that the extinction varies continuously over certain limited areas, which are separated from one another by discontinuous changes in the undulatory extinction. These discontinuities, in all her examples, are subparallel to the general orientation of the optic axes of the particular grain and impart to the grain, under crossed nicols, a banded appearance. Most of Hietanen's examples are highly metamorphosed, and the grains exhibit a banded appearance over the entire grain.

In the Baraboo quartzite, however, other discontinuities in the undulatory extinction are found. It is these discontinuities, and not the ones subparallel to the optic axis, that are termed "deformation bands." Adams[23] noted similar structures in vein quartz and described them as follows:

> The latticed structure shown by inclusions may sometimes be seen under crossed nicols (Pl. XXVI., *F*). The features brought out by this photograph are: (1) each single apparent veinlet consists of a thread of quartz which disagrees in extinction position with the parent crystal; (2) the crystallographic orientation of the quartz in parallel threads, the traces of roughly parallel planes, is the same; (3) the orientation in one set of planes is not the same as in the other set.

Deformation bands are infrequent in the Baraboo quartzite, but in two examples they are common (Samples 38 and 16, Figs. 9 and 10). The shift in the optic axis from grain to band is very small, so that the bands are observed only close to the extinction position of the grain. For fifty-four measured shifts in orientation from grain to band, the mean shift was 5°.9, with a standard deviation of 4°. The mode was in the 2°–4°-class interval, however, indicating that a few large values have substantially increased the mean. In addition, there is a suggestion that the distribution is bimodal. The surprising thing about the change in orientation is that in most examples the change is usually in the same rotational sense. In

[22] Pls. I–VII of ftn. 8.

[23] P. 648 of ftn. 9.

Sample 38, Figure 9*b*, for example, the sense of rotation from grain to band indicates that the younger beds to the south have moved upward with respect to the older beds to the north. In Sample 16, Figure 10*b*, however, the relationship is not clear cut. In many examples the difficulty is in determining which part of the grain is band and which part of the grain is the supposed undeformed part.

It is not possible to determine the relation between the deformation bands and the structure of the quartz, because the spatial orientation of the bands cannot be optically determined. The traces of the bands in the plane of the thin section, however, tend to be perpendicular and parallel to the optic axes. In addition, in Sample 38, Figure 9*b*, the traces of the deformation bands are mainly at a small angle to the bedding. If the bedding surface has acted as the shearing surface, then the relation of the traces of the deformation bands to the bedding is similar to the results of experimental work of W. Riedel, which has been discussed by A. Nádai.[24]

Riedel sheared plastic clay and found

[24] *Plasticity* (New York: McGraw-Hill Book Co., 1931), p. 330.

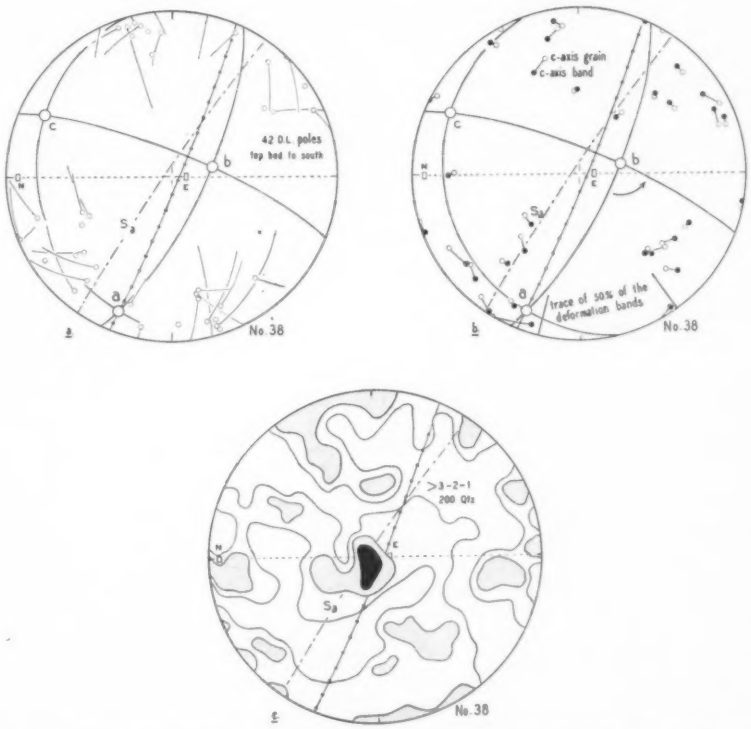


FIG. 9.—*a*, Pattern of poles of the deformation lamellae. The *a*–*b*-fabric plane is close to the axial-plane foliation determined in the field, from the hand sample, and in the thin section. *b*, Change in orientation of the optic axis from the major portion of a grain to the deformation band. All shifts are in the same rotational sense. *c*, Slight preferred orientation of the optic axes. The weak north-south girdle and the maximum at the center of the projection are reproducible.

that shearing cracks formed about 12° to the direction of shear. These are oriented similarly to the traces of the deformation bands in Sample 38, if the bedding plane is taken as the shear plane and the direction of shear is taken from the change in optic orientation from grain to band.

These bands need not be simply deformed quartz but could be a stress modification of SiO_2.

III. MICROFRACTURES

Microfractures are observed in most examples of the Baraboo quartzite. These fractures can be divided into those which are confined to individual crystals and those which pass through the aggregate of crystals without deviation at the crystal boundaries. It is the latter which exhibit a very strong preferred orientation and whose orientation pattern correlates very closely with the fabric axes determined from the pattern of the poles of the deformation lamellae. Sample 58, Figure 11, exhibits the close correlation of deformation-lamellae and microfracture patterns. The maximum of the poles of the microfractures lies but a few degrees from the *a*-fabric axis, determined independently from the deformation

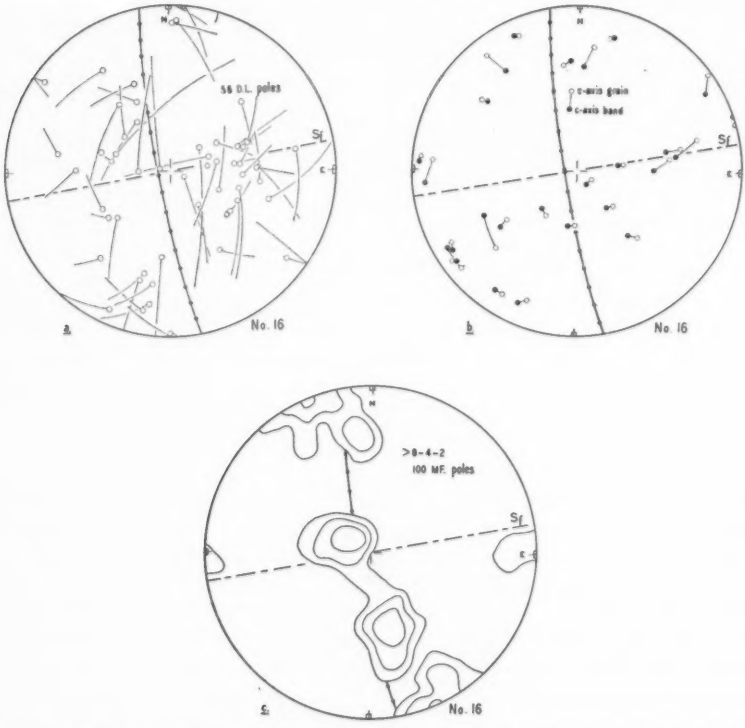


FIG. 10.—*a*, Diffuse pole pattern of deformation lamellae. The fabric axes for this sample shown in Fig. 14 are questionable. *b*, Change in orientation of the optic axis from grain to deformation band. Slight tendency for rotation to the southwest. The top of the bedding is to the east, and the younger beds have moved up with respect to the older. *c*, Poles of microfractures form a partial girdle, which is perpendicular to the shear fractures and subparallel to the bedding.

lamellae. A corresponding relation is shown in Sample 21, Figure 12. These two examples make it difficult to avoid the inference that the deformation lamellae and the microfractures are both related to the same set of external forces. In addition, since the deformation lamellae in the Baraboo quartzite are not consistently related to the bedding and the foliation, the microfractures are not so related.

The principal set of microfractures in the foregoing two examples and in all examples observed in the Baraboo quartzite are oriented very close to the *b*–*c*-fabric plane determined from the deformation lamellae. They are, therefore, termed, "*b*-*c* microfractures." Since the *a*-fabric axis is generally close to vertical, the *b*–*c*-fabric plane and the *b*-*c* microfractures are close to horizontal. In one sample, however—Sample 7, Figure 14—the *a*-fabric axis, determined from the deformation lamellae, lay close to the horizontal. The maxima of the poles of microfractures from this sample followed the above examples and coincided with the *a*-fabric axis.

In several patterns of poles of microfractures, two maxima are present. One

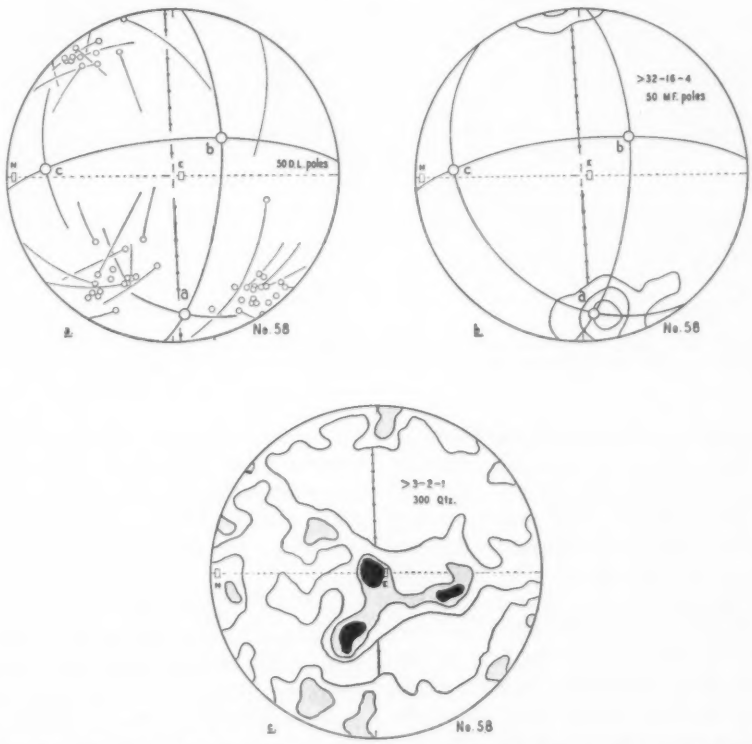


FIG. 11.—*a*, Symmetrical pole pattern of deformation lamellae, giving accurate location of the fabric axes. The sample is from a massive quartzite. *b*, The maximum of poles of microfractures is a few degrees from the *a*-fabric axis determined from the deformation lamellae. *c*, Weak north-south optic-axis girdle, partial horizontal girdle, and maxima close to axis of the weak north-south girdle.

maximum is close to the *a*-fabric axis, the other is weaker and close to the *b*-fabric axis. This type of pattern is shown in Sample 21, Figure 12, referred to above. These latter microfractures are termed "*a-c* microfractures." No examples were found in which *a-c* microfractures were present and *b-c* microfractures absent. In addition to the above two maxima, a few samples exhibit rare microfractures in the *a–b*-fabric plane.

Sample 16, Figure 10, evidences a partial girdle of poles to microfractures. O. F. Tuttle[25] in an abstract of his study pole girdle observed in Sample 16 is approximately east-west, corresponding closely to the strike of the region and perpendicular to the ever present east-west shear fractures. It was not possible to correlate the microfractures with the deformation lamellae in this sample because of the diffuse pattern of poles to the deformation lamellae.

The brief study of the microfractures in the Baraboo quartzite indicates that microfractures can be correlated with the pattern of the poles of the deformation lamellae and that *b-c* microfractures

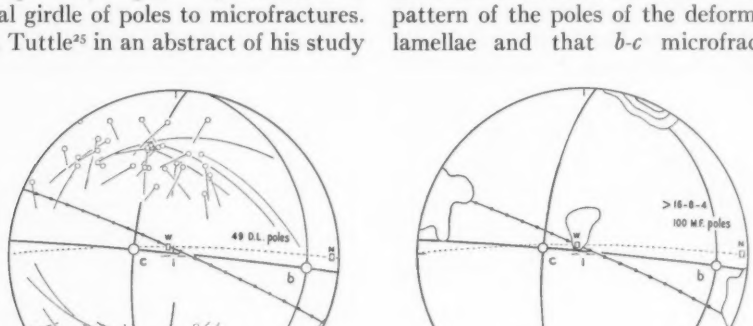


FIG. 12.—*a*, Symmetrical pole pattern of deformation lamellae. *b*, Three maxima of poles of microfractures at right angles. The principal maximum is close to the *a*-fabric axis determined from the deformation lamellae.

noted that in the rocks of the Washington area planes of inclusions passing through grain boundaries (termed "microfractures" in this paper) are not in agreement with local structure but are uniform over 15 or 20 miles. Also, he says that, in drag folds, planes of inclusions develop in shear directions and that one set predominates. Further, he observed that poles of the inclusion planes tend to be in a partial girdle about the fold axis. The axis of the microfracture

[25] "Structural Relations of Planes of Liquid Inclusions in the Crystalline Rocks of the Washington (D.C.) Area," *Amer. Geophys. Union Trans.*, Vol. XXVI (1945), p. 301.

are present if any microfractures are present. Since the pattern of the poles of the deformation lamellae do not correlate with the macroscopic foliation, the microfractures also do not so correlate.

IV. QUARTZ AND ZIRCON OPTIC-AXES FABRIC

The results of the study of the orientation of the quartz optic axes were discouraging. Many of the patterns are not reproducible from different parts of the same thin section. Three features of the weak patterns, however, do stand out and are usually reproducible: (1) a weak north-south vertical partial girdle; (2) a

small maximum close to the bedding and in the weak north-south girdle; (3) a small maximum close to the bedding and subperpendicular to the weak north-south girdle. In addition to the above features, it is possible in some examples to recognize a weak horizontal girdle.

The more definite examples of preferred orientation of the optic axes are found at the extreme eastern end of the outcrop area. Figure 5*b*, Sample 28, and Figure 13*a*, Sample 32, are two examples (Sample 32 is located near Sample 35, Fig. 14). No interpretation of any of the quartz-axes fabrics is attempted in this paper.

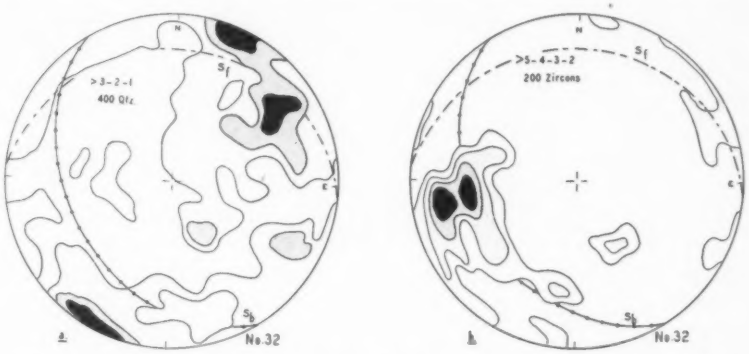


FIG. 13.—*a*, Pattern of optic axes of random quartz grains. A weak north-south girdle, dipping eastward, correlates with the pattern of Sample 28, Fig. 5*b*, which is on the south limb of the nose of the syncline. *b*, Pattern of optic axes of the zircon grains.

An examination of the outcrops, hand samples, and thin sections in the eastern half of the region shows that the deformation and the degree of preferred orientation increase from west to east toward the nose of the synclinorium. Comparisons among the samples can be made only among those which contain roughly the same admixture of sericitic material. This will be restated in the summary.

A flood of zircon grains was found in Sample 32, Figure 13*b*, located near Sample 35, Figure 14. The zircons occur in hematite laminae which are interbedded with laminae of quartz grains. The laminae are roughly 1 mm. thick. A fabric diagram was constructed for the optic axes of 200 zircons. One large maximum is located along the bedding plane. The individual zircons are fractured, but all exhibit rounded boundaries, while the quartz grains, a fraction of a millimeter away, rarely exhibit an original boundary and have been dimensionally elongated. The zircons are embedded in hematite (probably originally detrital iron oxide), and their orientation could have been accomplished by the dimensional orientation of the rigid zircons in a softer medium—i.e., hematite. This interpretation, of course, implies a correlation between the optic and the dimensional grain axes.

V. DIMENSIONAL GRAIN ELONGATION

The true spatial distribution of the axes of grain elongation is impossible to determine uniquely from thin-section studies. Lowe[26] determined a mean lineation from variously orientated outcrop surfaces of granite by assuming that the

[26] Ftn. 18.

grains are cylindrical in shape. In addition, he says that he has applied the same procedure to variously oriented thin sections of the same sample apparently by noting the modal direction in each thin section. These assumptions are definitely not true for most minerals developed through dynamic metamorphism; and, even if they were true, it is not possible to determine the actual frequency distribution from the frequency distribution of its projections. The above technique, however, does give an indication of the most frequent elongation if the grains are roughly cylindrical in shape and was applied to a few examples of the Baraboo quartzite.

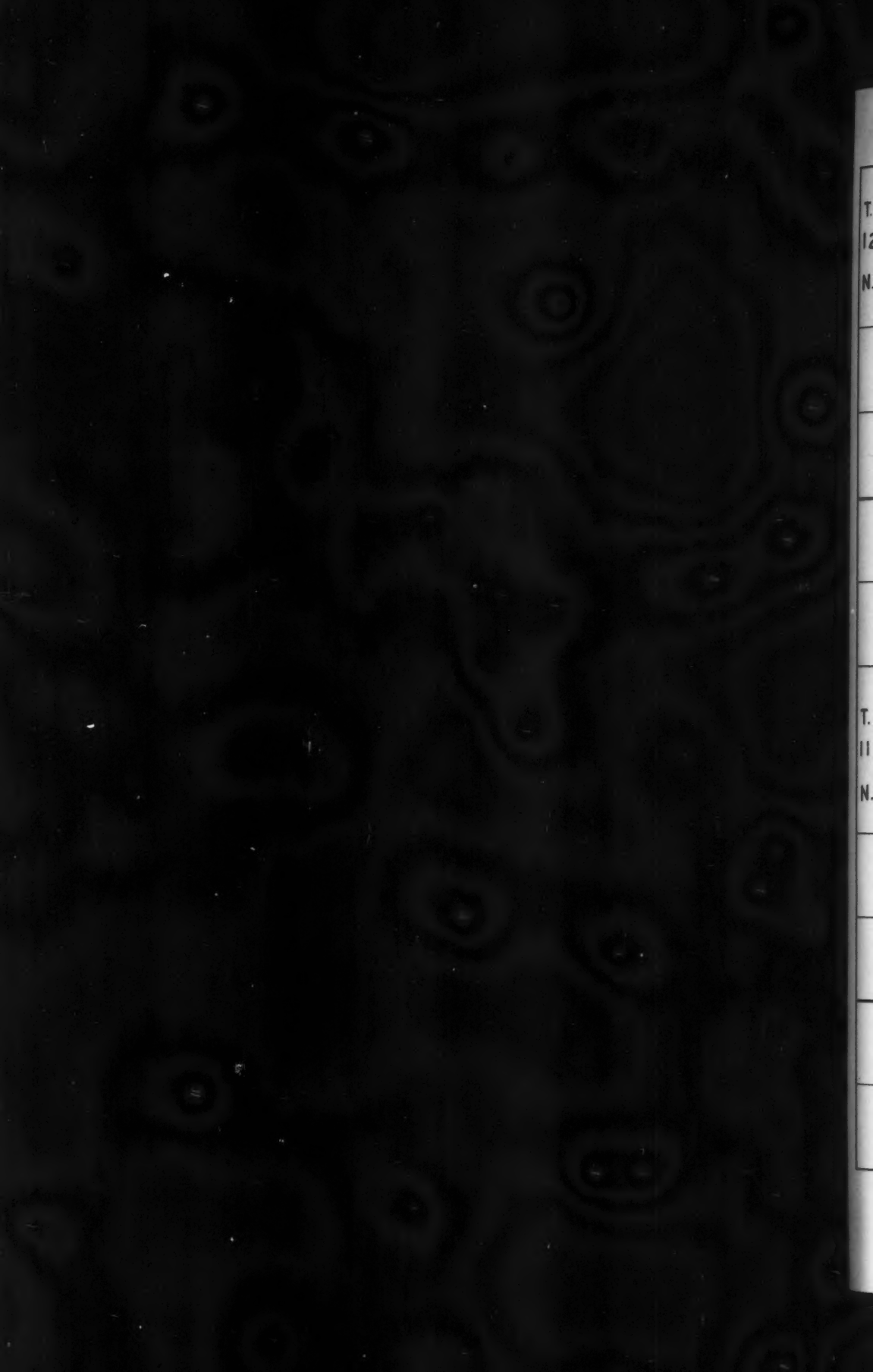
Only five clear-cut examples of grain elongation were encountered in the samples studied (Fig. 14). In all except one example, the major elongation of the grains trends in an east-northeast direction. This direction is also the approximate strike of the bedding and the axial-plane foliation. Most workers in structural petrology consider that the grain elongation is usually parallel to the intermediate strain axis of the deformation, that is, to the fold axis. If that is true in the Baraboo region, the grain elongation is the result of the major deformation, with a fold axis striking roughly east-northeast.

The relation between the grain elongation and the pattern of deformation lamellae has been mentioned before in connection with the deformation lamellae. The writer does not believe it feasible to determine the exact relations between deformation lamellae and elongation from the Baraboo region, because we are not sure of the number of deformations, the orientation of the forces causing the deformation, etc. A laboratory study and/or a field study in a region in which the boundary conditions are well known should indicate the solution.

SUMMARY

1. A study of Figure 14 indicates that the strike of the *a*–*b*-fabric plane, determined from the pattern of the deformation lamellae, varies over the synclinorium and, with very few exceptions, cuts across the strike of the bedding and foliation. If these fabric directions are found in other regions to correlate with the foliation, as has been reported by Ingerson and Tuttle, then the patterns of deformation lamellae in the Baraboo quartzite represent a deformation different from that which resulted in the macrofoliation.

On the other hand, the *a*-, *b*-, and *c*-fabric directions so determined could represent the local strain axes of the deformation, independent of the discontinuous mass movement along bedding surfaces. Strata containing axial-plane foliation were initially regions of weakness, and it is along these surfaces (for a wide range in the orientation of the applied forces) that the majority of the mass transport occurs. Referring again to Figure 14, we see that most of the *a*–*b*-fabric planes are subperpendicular or subparallel to the bedding and the foliation surfaces. If we think of the whole structure as having been formed in one deformation, the regions in which the *a*-*b* fabric is subperpendicular to the bedding and foliation are regions of approximately horizontal compression parallel to the bedding and foliation. Conversely, regions in which the *a*–*b*-fabric plane is subparallel to the bedding and the foliation are regions of approximately horizontal stretching parallel to the bedding and the foliation. This latter example is the usual concept of the orientation of the strain in a simple fold. The above-described regions of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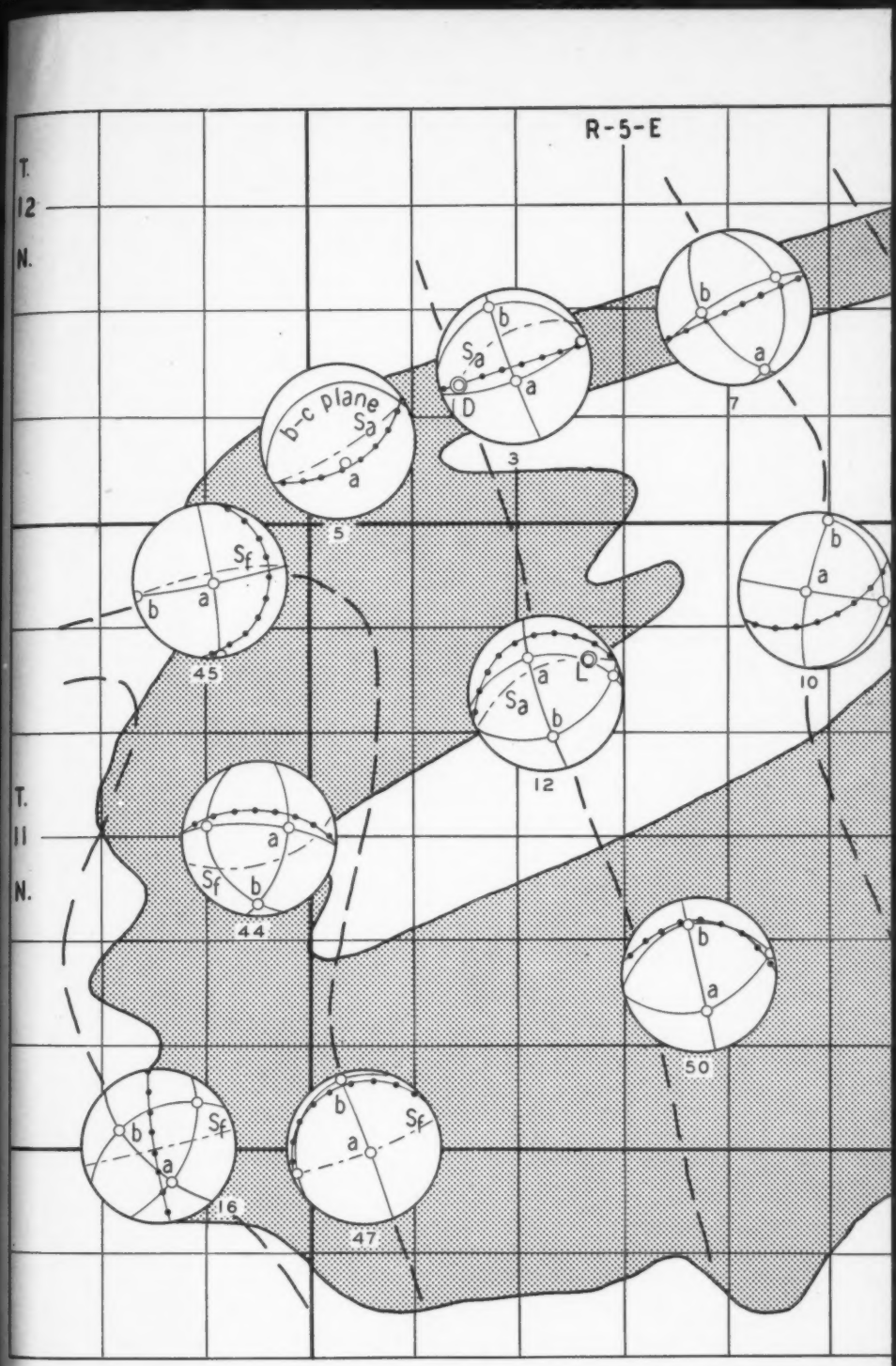
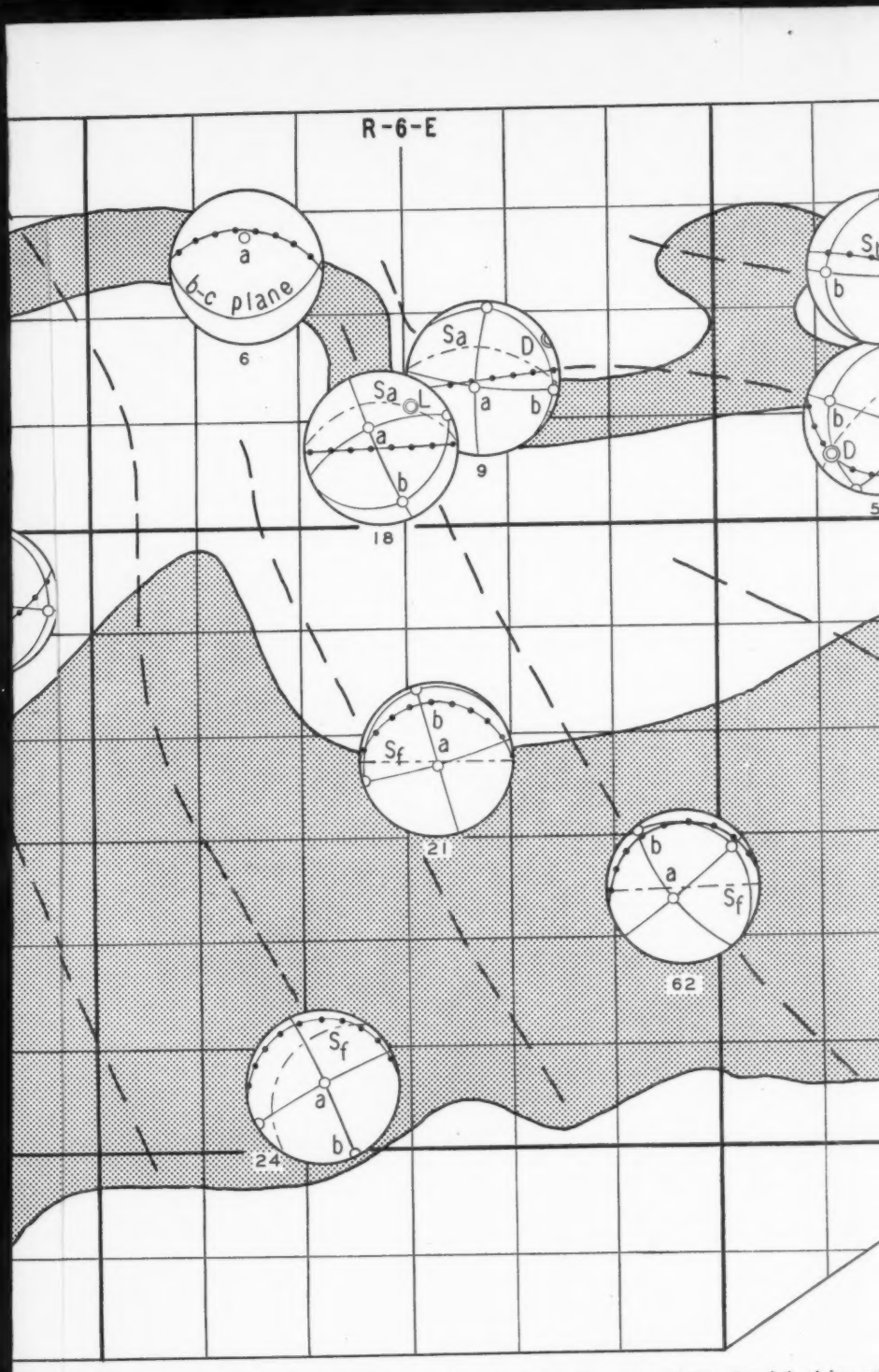
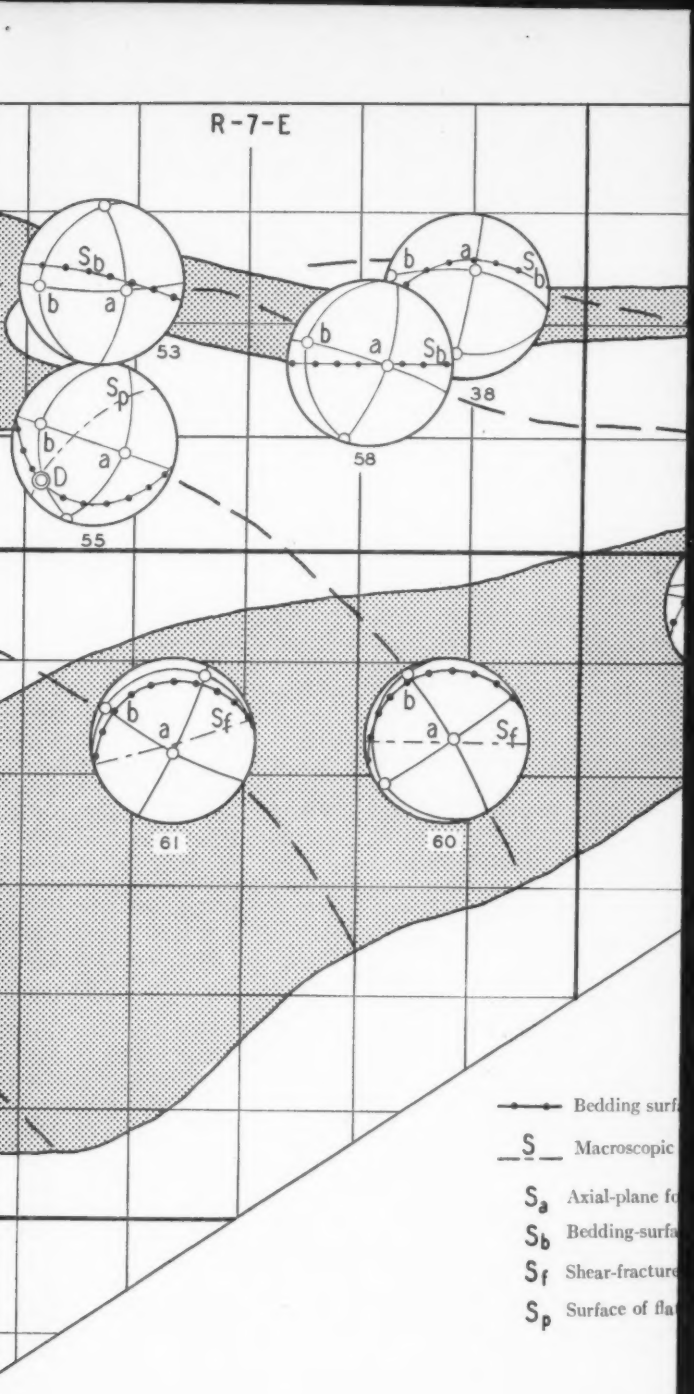


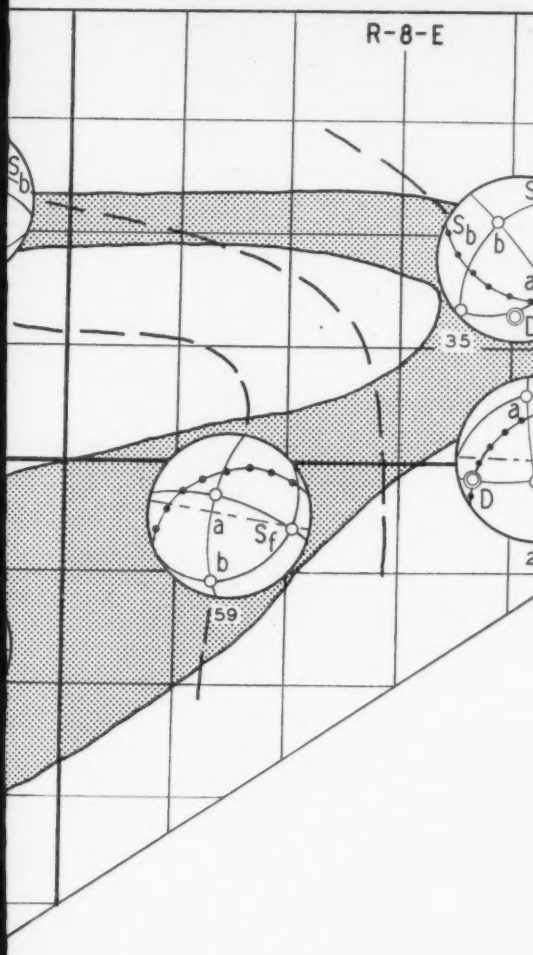
FIG. 14.—Generalized outcrop area of the Baraboo quartzite (modified



odified from A. Leith). The *a*-, *b*-, and *c*-fabric axes were determined from the pattern of the poles of the deformat



leformation lamellae. The dashed lines represent the trend of the strike of the *a*–*b*-fa



Bedding surface

Macroscopic foliation

S_a Axial-plane foliation

S_b Bedding-surface foliation

S_f Shear-fracture surface

S_p Surface of flattened pebbles

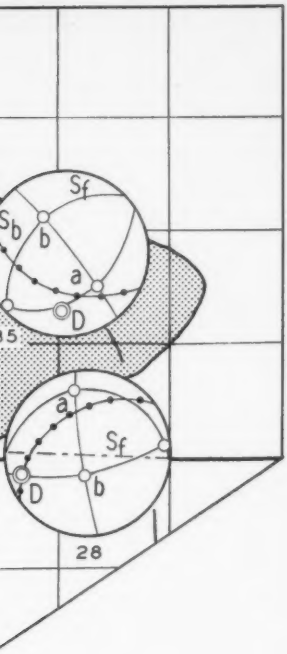
D Mean direction elongation

L Axis of "buckling

Projection of the

N D, E D Compass direction

strike of the *a*–*b*-fabric plane so determined



direction of dimensional grain
ngation

of "buckling" of foliation surface

ction of the horizontal

ass directions

compression and stretching in the bedding and/or foliation plane can be thought of as regions of convergence and divergence perpendicular to the principal direction of movement. This phenomenon is recognized in ocean currents and atmospheric currents. Koenigsberger and Hungerer,[27] studying both elastic and plastic bodies, found that in the bending of prisms a secondary fold develops, with a larger radius of curvature than the primary fold and at right angles to the primary fold. These secondary folds have long been recognized by the structural geologists and are called "saddle folds" or "cross-folds." The orientation of the fabric axes determined from the deformation lamellae, therefore, could possibly have resulted from the secondary stresses produced from a series of initially east-northeast folds.

The writer has no preference for either of the above two interpretations. The field problem is always tremendously complicated, and it is almost pure fantasy to interpret a complicated structure in simple terms when we are not even able to interpret a simple structure in complicated terms.

2. In addition to the pattern of the fabric axes determined from the deformation lamellae, the following observations point toward an interpretation of structure that would necessitate more than one period of deformation:

a) Two or more periods of quartz veining are separated by at least one period of deformation.

b) Shear-joint surfaces or perhaps shear fractures in the western half of the region tend to be subparallel to the strike of the *a*–*b*-fabric plane determined from the deformation lamellae.

c) There is chevron folding, that is, buckling, of the axial-plane foliation.

3. A study of a series of samples from the eastern half of the region (in particular, from Sample 53 eastward, in the north range) indicates that the intensity of the deformation has increased toward the nose of the synclinorium. This is evidenced by the following observations:

a) Decrease in the percentage of grains showing original boundaries

b) Increase in the total angular variation in the position of the optic axis in individual grains

c) Increase in the number of grains showing discontinuities in the undulatory extinction

d) Increase in the number of grains which have been broken down into smaller units

e) Increase in the degree of preferred orientation of the optic axes

It is again emphasized that comparisons among samples can be made only when all the samples are of the same composition, that is, in this example, the amount of admixed sericite relative to the quartz. Sericite is a good lubricant, and, when admixed in sufficient quantities to the quartz grains, it will account for the majority of the change in shape of the mixture during deformation.

4. This study has convinced the author that structural petrology needs a theoretical foundation based on experimental deformation of solids and consistent with the physical and chemical concepts of the structure of matter.

[27] A discussion of their work is given by H. W. Fairbairn, *Structural Petrology of Deformed Rocks* (Cambridge, Mass.: Addison-Wesley Press, 1942), pp. 78 and 96.

STRUCTURAL PETROLOGY OF THE PINE CREEK AREA DICKINSON COUNTY, MICHIGAN

JAMES W. HIGGINS
Standard Oil Company of California, Los Angeles

ABSTRACT

The structures of Archean granite and gneiss and overlying Lower Huronian sedimentary formations in the vicinity of Pine Creek in the northern part of the Menominee iron-bearing district of Michigan are compared in this study. Foliation of the gneiss is distinctly unconformable with bedding of the sediments. Faults, joints, and microstructures are similarly developed in the two rock types. During the Killarney orogeny these Archean and Lower Huronian rocks were upturned into a southwest-facing monocline, pushed back at its ends along northeast-trending tear faults. The petrofabrics of both the granitic and the sedimentary rocks plot into horizontal *A–C* girdles and are presumably the result of rotation about a vertical axis associated with the faulting.

INTRODUCTION

The Pine Creek area of southern Dickinson County, Michigan, constitutes the northern portion of the Menominee iron-bearing district. The area occupies a belt 16 miles long and 1–2 miles wide, which trends northwest, generally adjacent to the course of Pine Creek. Included in this belt is the southwestern margin of a complex of granites and gneisses and an almost continuous strip of strongly metamorphosed sedimentary rocks that flank the complex.

In 1904 W. S. Bayley[1] placed these sedimentary rocks in the Lower Huronian and cited occurrences of their basal conglomerates truncating foliation, pegmatite veins, and other structures of the granitic rocks of the Pine Creek area as conclusive evidence that the granites are of Archean age. His opinion remained undisputed in print until C. A. Lamey, discussing the possibility of an extension of the post-Huronian Republic granite, expressed doubt as to the age of the conglomerate and stated that the conglomerate proved "merely that some granite was exposed at the time of its formation, but does not prove the age of that granite or whether its occurrence was widespread or local."[2] F. J. Pettijohn[3] has recently restudied the basal conglomerate and has defended the view of Bayley.

This controversy supplied the impetus for the present study. It seemed probable that a detailed examination and a comparison of the structures, especially the petrofabrics, of the adjacent granitic and sedimentary rocks would give decisive information concerning the relative ages of the rock systems. The author, therefore, spent twelve weeks of the summer of 1946 in the Pine Creek area.

GENERAL GEOLOGY

The Archean complex is composed principally of pink gneisses and granites; subordinate to these are gray gneiss, migmatite, and brick-red granite. All phases of the complex are widespread except the red granite, which forms only 5 per cent of the complex and is restricted to small scattered patches in the northwestern part of the area.

[1] "The Menominee Iron-bearing District of Michigan," *U.S. Geol. Surv. Mono. 46* (1904).

[2] "Republic Granite or Basement Complex?" *Jour. Geol.*, Vol. XLV (1937), p. 507.

[3] "Basal Huronian Conglomerates of Menominee and Calumet Districts, Michigan," *Jour. Geol.*, Vol. LI (1943), pp. 387–97.

The gneisses of the Archean complex are of primary igneous origin. Their folia are due to subparallel arrangement of layers of different composition. These folia vary locally in strike and dip; they wrap around xenoliths; and they wedge out and split into several bands. Much of the pink gneiss contains augen-like phenocrysts of potash feldspar, oriented parallel to the trends of folia, and in many outcrops this foliated pink gneiss grades imperceptibly into lineated pencil gneiss.

The upturned Lower Huronian sediments rest unconformably against the Archean complex. The basal part of the Lower Huronian series, termed the "Fern Creek formation" by F. J. Pettijohn,[4] is composed of 300 feet of conglomerate, arkose, graywacke, argillite, and phyllite. The conglomerate, which truncates structures of the Archean complex, was seen at the following localities: (1) at the falls of the Sturgeon River, Section 8, T. 39 N., R. 28 W.; (2) in the northwestern part of Section 8, T. 39 N., R. 28 W.; (3) at the falls of Black Creek, north-central part of Section 7, T. 39 N., R. 28 W.; (4) near Fern Creek, Section 34, T. 40 N., R. 29 W.; and (5) in the south-central part of Section 29, T. 41 N., R. 29 W.

Above the Fern Creek beds is the Sturgeon formation, which is divisible into a lower member of green sericitic quartzite and an upper member of white, vitreous, fine-grained quartzite. The average thickness of the Sturgeon formation is 1,000 feet. This formation grades into Randville dolomite, the upper member of the Lower Huronian system.

The youngest pre-Cambrian rocks of the area are a series of dikes and mineralized quartz veins, which intrude the Archean complex, and the Fern Creek and Sturgeon formations. The dikes are predominately metamorphosed gabbros and diabases, but in the northwestern part of the area occurrences of melasilexite and esmeraldite are noted.

[4] P. 397 of ftn. 3.

PRIMARY STRUCTURAL UNCONFORMITY

Although the existence of a major angular unconformity between the Archean complex and the Lower Huronian formations is proved by contacts of the basal conglomerates with the granitic rocks at well-spaced localities, this relationship is emphasized by comparing the general strike of the Sturgeon quartzite with the general trends of the foliation of the gneiss throughout the area.

The strike of the Sturgeon quartzite, consistently northwest, actually deviates between N. 15° W. and N. 85° W., and averages N. 50° W. In the eastern part of the Pine Creek area the foliation planes of the gneiss trend unvaryingly northwest and form an acute angle with the strike of the quartzite. West of Fern Creek, Section 34, T. 40 N., R. 29 W., the trends gradually change to north and even northeast, the angle between the foliation of the gneiss and the strike of the quartzite increasing to approximately 90°.

The discordance of trends of the quartzite and the gneisses is graphically illustrated by Figure 1, a diagram of the poles to bedding planes of the quartzite, and by Figures 2 and 3, diagrams of poles to foliation planes of the gneisses in the eastern and western parts of the area, respectively.

If the marginal granite gneisses were post-Huronian, their foliation, whether resulting from granitization of sediments or marking lines of flow of an intrusive massive, normally would conform to the general trend of the Lower Huronian rocks. Therefore, the hypothesis of post-

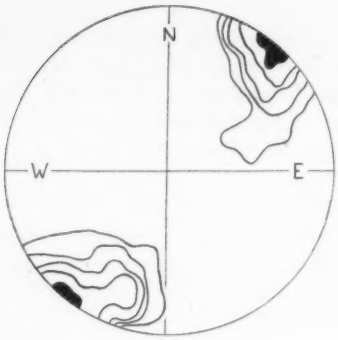


FIG. 1

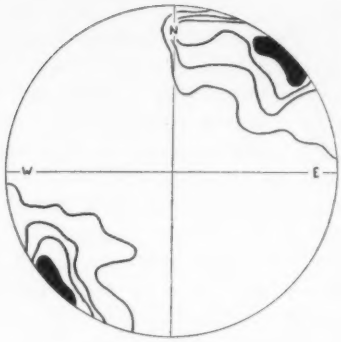


FIG. 2

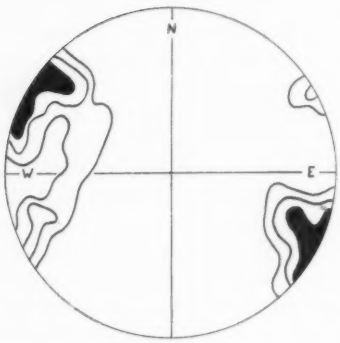


FIG. 3

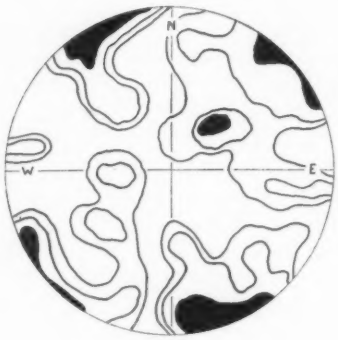


FIG. 4

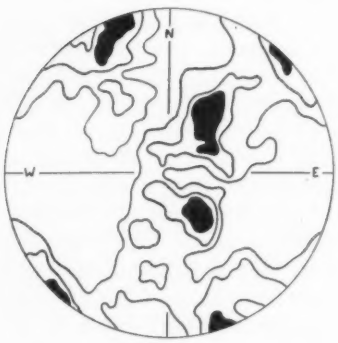


FIG. 5

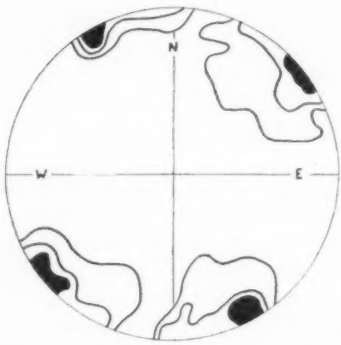


FIG. 6

FIGS. 1–6.—Fig. 1, poles of 119 bedding planes of quartzite west of Section 28, T. 40 N., R. 29 W. (20 per cent maximum). Fig. 2, poles of 50 foliation planes of gneiss east of Section 35, T. 40 N., R. 29 W. (12 per cent maximum). Fig. 3, poles of 70 foliation planes of gneiss west of Section 35, T. 40 N., R. 29 W. (10 per cent maximum). Fig. 4, poles of 76 joints of granitic rocks (4–7 per cent maxima). Fig. 5, poles of 301 joints of quartzite (4 per cent maxima). Fig. 6, poles of 36 joints of basic dikes (10 per cent maxima). All figures are plotted on the lower hemisphere of an equal-area net.

Huronian age of the granite complex is incompatible with the facts.

MAJOR DEFORMATIONAL STRUCTURES

The Lower Huronian strata of the Pine Creek area have been folded and subsequently faulted since their deposition. Both folding and faulting probably were separate phases of the Killarney orogeny.

During the folding the rocks were upturned into a monocline that faced southwest. Dislocated remnants of this monocline are preserved as bluffs of Sturgeon quartzite, which form an arc trending northwest from the center of Section 9, T. 39 N., R. 28 W., to the southwest corner of Section 20, T. 41 N., R. 29 W. The convexity of this line of bluffs is not due to progressive change of strike of the quartzite, however, but to progressive displacement of the quartzite along northeast-trending vertical faults.

The strike of the quartzite, as stated in an earlier paragraph, deviates between N. 15° W. and N. 85° W. and is generally N. 50° W. (Fig. 1). The deviations are local and sporadic and are caused by contortion and rotation of faulted blocks. The dip of the quartzite also has been markedly affected by fault-block rotation. Where faulting is not conspicuous, the beds are vertical or dip steeply southwest. At many places where faulting has been forceful, the beds are overturned and dip northeast, evidently the consequence of tilting of faulted blocks. However, some of the overturning may be the result of underthrusting of quartzite during folding.

Faults of the Pine Creek area are of the tear type. Twenty-one of these faults can be safely inferred and are shown on the structure map, but other faults of inappreciable displacement may exist. Fifteen of these faults are concentrated in the northwestern Pine Creek area. All fifteen strike northeast, their average trend being N. 58° E. Along eleven of the faults the northwest blocks have been displaced northeast through an aggregate distance of more than 3.5 miles. A relatively southwest displacement of northwest blocks along the other four faults totals approximately 1,100 feet.

If the displacements along the fifteen faults are added vectorally and the resultant resolved into east-west and north-south components, the total displacement is 3.5 miles in a direction N. 70° E., the east-west component is 3.2 miles, and the north-south component is 1.2 miles (Fig. 7). The fact that the east-west component has nearly three times the magnitude of the north-south component implies that the forces of faulting were essentially directed from the west.

Related to the faulting of the northwestern area are shear zones strongly developed in both the quartzite and the granite. The rocks in these zones are brecciated or cleaved along closely spaced planes. Some of the zones are found in the immediate vicinities of mappable faults. However, many occur at considerable distances from definite faults, but may themselves be minor faults along which displacement is not prominent.

All except three of seventeen observed shear zones of the quartzite trend between N. 55° E. and east-west. The average trend of all the zones is N. 67° E., differing from the average strike of the faults (N. 58° E.) by only 9° and from the trend of the resultant of the fault movements (N. 70° E.) by only 3°. In several instances shear planes are strongly striated. The pitch of these striations ranges from nearly horizontal (2° NE.) to 40° SW. The pitch of the striations and the average strike of the shear zones

are evidence that the general shear movement was upward at a low angle from west of southwest.

The trends of nineteen observed shear zones in the granite range between N. 15° E. and east-west, and average N. 52° E. This latter figure differs from the average trend of the faults by only 6° and from the average trend of the shear zones in the quartzite by 15°.

Brick-red granite, the only phase of the granitic complex which is nowhere truncated by Lower Huronian basal conglomerate, is intensely sheared in several outcrops. Thus this granite is proved to be older than the faulting, although not necessarily Archean.

At the southeastern end of the Pine Creek area three faults have been mapped. These faults have strikes of N. 44° E., N. 45° E., and N. 60° E., and horizontal displacements of 300, 100, and 100 feet, respectively. In each instance the southeast block has been displaced northeast, the relative movement being opposite to that in the northwestern part of the area.

The central part of the area is but slightly faulted, only three small northeast-trending faults in the vicinity of Fern Creek, Section 34, T. 40 N., R. 29 W., being observed in a distance of 7 miles. Moreover, the quartzite of this area is characteristically sheared parallel to bedding planes. In some outcrops the long dimensions of autoclasts of quartzite are oriented subparallel to the strike of the beds. Therefore, it is assumed that the shearing has resulted from lateral, rather than from vertical, slippage of the bedding planes.

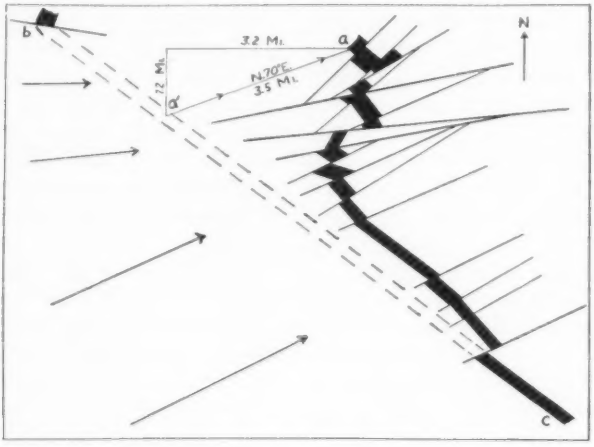


FIG. 7

The faults of the Pine Creek area are not local phenomena but are elements of a regional system. C. E. Dutton and C. A. Lamey have found similar faults in parts of the Menominee iron-bearing district which lie several miles south of the granitic complex. They write:

> Likewise, Although it is recognized that folding is an important factor in the regional distribution of formations, faulting evidently dominates their local distribution. Furthermore, a systematic orientation of faults is apparent, and a well-defined fault pattern comprising (1) a conspicuous set of faults with northeast-south-

west trends, and (2) less conspicuous sets trending (*a*) northwest-southeast, (*b*) nearly north-south, and (*c*) approximately east-west.[5]

The concentration of faults in the northwestern and southeastern parts of the Pine Creek area and the characteristic development of bedding-plane shear in the least faulted central part of the area are best explained by assuming that the Pine Creek monoclinal structure was reinforced by the massive complex of Archean granite. Forces pushing from the southwest against the buttressed central part of the monocline were deflected parallel to the monoclinal front. Although these forces were unable to fault the unyielding granite massive at its center, their deflection was accompanied by lateral slippage of the less competent quartzite along bedding planes. On the other hand, the same forces faulted and pushed back the monoclinal structure at its least resistant ends. The total movement is analogous to that of the front of a glacier thrusting against and moving around a nunatak.

If successive fault blocks in the northwestern part of the Pine Creek area are moved back along their planes, beginning with the fault of Section 20, T. 40 N., R. 29 W., distances equal to their displacements, an approximation of the original front of the Pine Creek monocline can be obtained. The northwesternmost outcrop of quartzite now in the southwest corner of Section 20, T. 41 N., R. 29 W., is moved to a position in the northwest corner of Section 35, T. 41 N., R. 30 W. The same position is obtained if the northwesternmost outcrop is moved a distance equal to, and in a direction opposite to, its total displacement, that is, 3.5 miles in the direction S. 70° W. This position is on the projected strike (N. 53° W.) of unfaulted quartzite, which crops out in Section 28, T. 40 N., R. 29 W., and is less than 3 miles from Sturgeon quartzite near Browning Creek, Section 20, T. 41 N., R. 30 W., which has been described by Pettijohn.[6]

These relationships are pictured in Figure 7. The present and reconstructed positions of the northwesternmost outcrop of quartzite in the Pine Creek area are represented by *a* and *a'*, respectively. The quartzite at Browning Creek is *b*, and the unfaulted quartzite is *c*.

Although the present distance between the Pine Creek and Browning Creek exposures of quartzite is about 6 miles, it is thought that, before faulting, the quartzites at these localities were connected parts of the Pine Creek monocline. This monocline evidently curved northward in the vicinity of Browning Creek, because Pettijohn has determined the average strike of the quartzite at that locality to be N. 35° E.[7]

Two structural features of the Archean complex have resulted from the deformation during the folding of the Pine Creek monocline. The first of these is boudinage structure, occurring in several outcrops in the northwestern part of the area, where lenses of granitic material, 6 inches to 3 feet thick, have been injected into what are now biotite-hornblende schists. These competent lenses dip toward the quartzite and have been broken along their dips into links or boudins surrounded by the more plastic schists. The development of these boudinage structures is attributed to stretching of the granitic lenses as they yielded to forces which folded the quartzite.

A second evidence of deformation of the Archean gneiss is a thrust fault zone

[5] "Geology of the Menominee Range Vicinity of Iron Mountain," *Mich. Dept. Conserv., Geol. Surv. Div. Prog. Rept. 6* (1941), p. 3.

[6] Pp. 395–97 of ftn. 3.

[7] *Ibid.*, p. 395.

in the eastern part of the area. This zone, about ¼ mile north of the margin of the complex, is marked by a sharp reversal in dip of foliation planes of the gneiss and is traceable from Section 9, T. 39 N., R. 28 W., to Section 27, T. 40 N., R. 29 W., a distance of 6 miles (see Map I). Folia of the gneisses in outcrops south of this zone strike unvaryingly northwest and dip southwest, whereas the folia in outcrops north of the zone also strike consistently northwest but dip northeast. Furthermore, northwest strike and northeast dip are characteristic of gneisses throughout the width of the complex north of this zone of dip reversal. Because the northwestern part of the Pine Creek area is so complexly faulted, no integrated zone of dip reversal could be determined there.

Inasmuch as the southwest-dipping gneisses of the margin of the complex, that is, south of the zone of dip reversal, are in direct contact with, and truncated by, the basal conglomerate of the sedimentary series, it is evident that the gneisses at these contacts have been upturned through the same angular distance as the sedimentary rocks. Therefore, the attitudes of the gneisses before deposition of the sediments can be obtained simply by solving the "two-tilt" problem on the stereographic net.[8] This solution involves merely the rotation of a particular plane of foliation about an axis corresponding to the strike of adjacent quartzite and through an angle of such magnitude that the quartzite becomes horizontal.

Folia of gneisses of ten well-spaced outcrops adjacent to quartzite have been rotated in this manner, and their resulting attitudes have been plotted on a map (Fig. 8). To this map have been added typical attitudes of nonrotated folia north of the zone of dip reversal. Finally, the approximate original locations of all folia have been obtained by reconstructing the faulted blocks. The result is a concatenated pattern suggestive of a pluton, the center of which probably was near Fern Creek, Section 34, T. 40 N., R. 29 W., where trends of the foliation change from northwest to northeast.

Therefore, an assumption that the marginal gneisses originally had trends similar to those obtained by rotation, but were subsequently overturned, probably during the formation of the Pine Creek monocline, is indeed credible. Inasmuch as the zone which separates the overturned and nonfolded gneisses is so sharply delimited, it may reasonably be considered a fault zone. This fault probably formed along the axial plane of the monocline and most likely was developed by thrust from the northeast. Such thrusting would have helped to overturn the marginal gneisses.

JOINTING

Similar systems of joints in the granite gneisses and the Sturgeon quartzite are considered evidence that the Archean and Lower Huronian rocks have the same deformational history. The similarity of the joint systems of these rocks is demonstrated by Figures 4 and 5.

One system of joints in the granite consists of a set that strikes N. 45° W. and dips 45° SW., and a weaker set that strikes N. 45° W. and dips about 40° NE. A system of joints in the quartzite includes a set which strikes northwest and dips 45° SW., and a set which strikes northeast and dips 30° NW. If these joints in the granite and quartzite are assumed to be compressional, the bisector of the acute angle of the two sets in

[8] W. H. Bucher, "The Stereographic Projection: A Handy Tool for the Practical Geologist," *Jour. Geol.*, Vol. LII (1944), pp. 203–4.

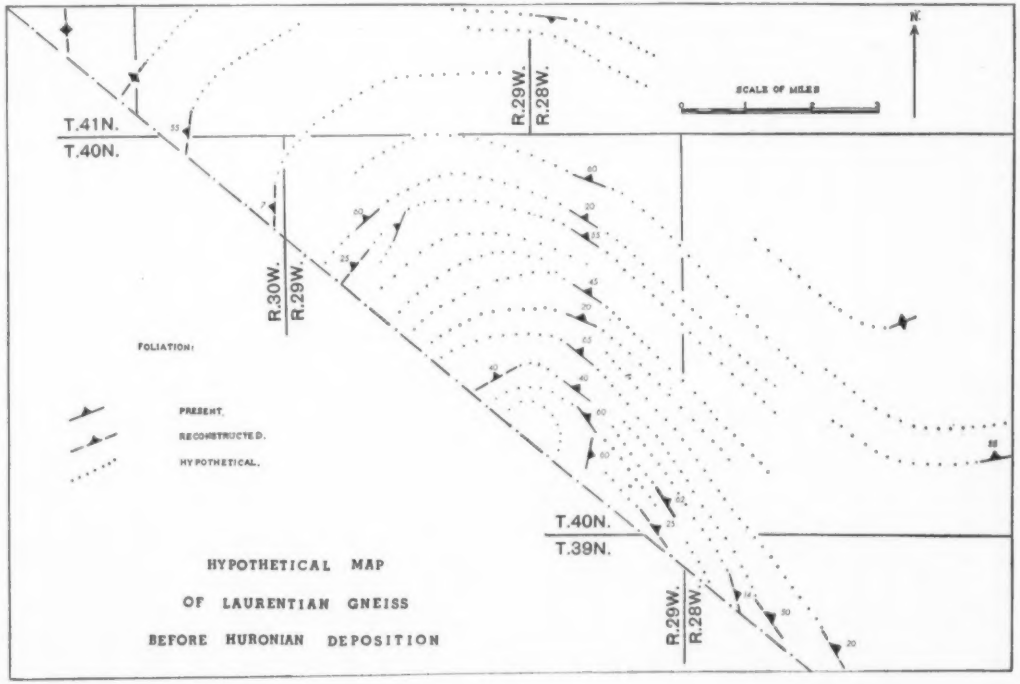

T.41N.
T.40N.
R.30W.
R.29W.
R.29W.
R.28W.
SCALE OF MILES
N.
FOLIATION:
PRESENT.
RECONSTRUCTED.
HYPOTHETICAL.
T.40N.
T.39N.
R.29W.
R.28W.
HYPOTHETICAL MAP
OF LAURENTIAN GNEISS
BEFORE HURONIAN DEPOSITION

FIG. 8

each rock type is equal to the direction of maximum compression. This direction, obtained by use of the stereographic net as outlined by W. H. Bucher,[9] is 15° upward from S. 10°W. if determined by the joints in the quartzite and is nearly horizontal from S. 40° W. if determined by joints in the granite. These directions correspond closely enough to postulate

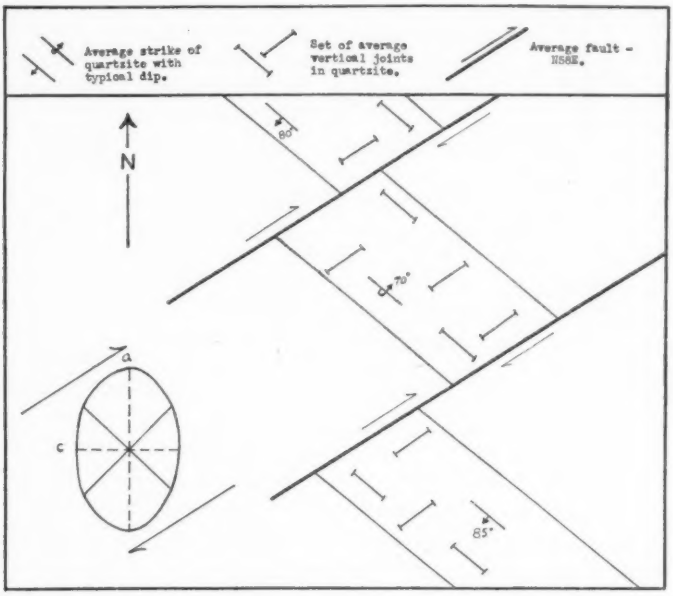


FIG. 9

that the conjugate sets of joints of low-angled dip in the quartzite and granite probably resulted from the same forces of deformation. The general direction of the compression which produced these systems of joints is approximately perpendicular to the trend of the Pine Creek monocline. Therefore, the joints may have formed during the folding of this monocline.

A second system of joints, consisting of two vertical sets which trend northeast and northwest, is also equally present in both the granite and the quartzite. The northeast-trending joints, the most conspicuous of the Pine Creek area, locally are the controlling factors of topography. The northwest-trending joints have an average attitude similar to that of the bedding of the quartzite, and many bedding planes are separated by these joints. However, in localities where the bedding differs in strike and dip from the average, these northwest-trending vertical joints are quite distinct from bedding planes.

Both sets of the vertical joints are particularly well shown near zones of faulting, especially in the northwestern part of the area. They may be shear joints related to a couple produced by differential movement of the fault blocks. Figure 9 discloses the possible relationship of these joints to the faulting in the north-

[9] *Ibid.*, p. 196.

western part of the area and to the strain ellipsoid. The strain ellipsoid of the figure indicates that the couple produced an effective component of elongation (*a*) in a north-south direction, and one of compression (*c*) in an east-west direction. The directions of these components of force agree with those obtained by an earlier vectoral analysis of the fault movements of the northwestern area.

The joints of post-Huronian basic dikes of the Pine Creek area belong to a system of vertical sets which correspond in strike to the vertical sets of the granite and quartzite (Fig. 6). Apparently, therefore, the joints of the basic dikes were produced during the faulting. On the other hand, the absence in the dikes of the sets of low-angled joints, common to both the quartzite and the granite gneiss and supposedly related to the folding, is evidence that the dike rocks were intruded either during final stages of the folding or later. Thus the rocks of the granitic complex are not only clearly divorced from, but also definitely older than, the only major features of post-Huronian igneous intrusion. This fact emphasizes the probability of Archean age of this complex.

PETROFABRICS

To supplement the structural data obtained in the field, a petrofabric study has been made of eight samples of the granite and twelve samples of the quartzite. The positions of these samples relative to major structures can be found on the structure map (Map I). The samples of granite and gneiss are designated by the letters A through H, and the quartzite by numerals 1 through 12.

It is not considered necessary to give an elaborate discussion of the theory and procedures of petrofabric analysis. These subjects are treated thoroughly in books by Bruno Sander,[10] Eleanor Knopf and Earl Ingerson,[11] H. W. Fairbairn,[12] M. P. Billings,[13] R. C. Emmons,[14] and in excellent papers by Anna Hietanen[15] and by R. E. Fellows.[16]

METHODS

All samples were oriented in the field relative to some *S*-plane of determined strike and dip, usually a bedding or foliation plane, less frequently a shear plane. Two sections of each specimen were cut for study, one perpendicular to the strike and parallel to the dip of an *S*-plane, the other at right angles to the first, perpendicular to the dip and parallel to the strike.

A study was made of the orientation of the *c*-axes of 200–300 quartz grains in each thin section. For this study a petrographic microscope of the Leitz type and a 4-axis Universal Stage were used. The size of the quartz grains in all samples ranges from less than 0.05 mm. in crush-quartz to more than 1.0 mm. in original grains. Only larger grains were chosen for measurement, as the importance of discrimination in grain selection is becoming increasingly recognized.[17] However, all

[10] *Gefügekunde der Gesteine* (Vienna: Julius Springer, 1930).

[11] "Structural Petrology," *Geol. Soc. Amer. Mem. 6* (1938).

[12] *Structural Petrology of Deformed Rocks* (Cambridge: Addison-Wesley Press, Inc., 1942).

[13] *Structural Geology* (New York: Prentice-Hall, Inc., 1942), pp. 352–55.

[14] "The Universal Stage," *Geol. Soc. Amer. Mem. 8* (1943).

[15] *On the Petrology of Finnish Quartzites* (Helsinki: Government Press, 1938).

[16] "Recrystallization and Flowage in Appalachian Quartzite," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1399–1432.

[17] *Ibid.*, p. 1422; also "The Influence of Grain-Selection on the Meaning of Quartz Diagrams," *Amer. Geophys. Union Trans.* (1945), pp. 653–59.

the large grains were selected at random along five or six well-spaced traverses across the slide. Two of the diagrams are more selective than the others because only those grains showing some degree of undulatory extinction were chosen.

To insure the existence of homogeneity, the grains were plotted in separate groups of 50 or 100, and an inspection made for the tendencies of concentration after each group had been plotted. The plotting was done on the lower hemisphere of an equal-area net of the so-called "Schmidt" type. Concentrations of the *c*-axes were determined by standard center and peripheral counters of 1.0-cm. radius. Counting was spaced by grid distances of 1.0 cm. This was supplemented by closer counting in areas of maxima.

The diagrams which have resulted from the study of sections cut perpendicular to *S*-planes have been rotated into a horizontal plane and are plotted on the structure map. The procedure of rotation of these diagrams was that outlined by Ingerson.[18]

OBSERVATIONS AND INTERPRETATION

The most apparent feature of the fabric diagrams is that they all have peripheral maxima. On the basis of position of the peripheral maxima, the diagrams may be divided into three types.

One type of diagram is characterized by maxima lying between east-west and northeast-southwest, being transverse to any primary structures of the rocks. Diagrams of this type are restricted to samples collected from the southeastern Pine Creek area.

A second distinct type is most common to samples of the northwestern part of the area. Maxima of this type have positions generally north-south. The orientations of these maxima are transverse to bedding planes of the quartzite but usually subparallel to foliation of the gneiss.

To a third type belong a remaining few diagrams of quartzite. The maxima of this type have positions subparallel to bedding planes of the quartzite and may be related merely to residual sedimentary fabric, but more probably to bedding-plane shear. The diagrams of Samples 3, 4, and 5 of the quartzite belong to this type.

Many of the maxima of the diagrams are point maxima, but a few, such as those of Samples 2, 4, and 6 of the sedimentary rocks and Samples A and E of the granite, belong to horizontal girdles. If these horizontal girdles are *A–C* girdles, the directed force must have acted in a nearly horizontal plane about a vertical axis of rotation. Such a force was associated with the faulting. Therefore, a comparison of the fabrics and the faults is desirable.

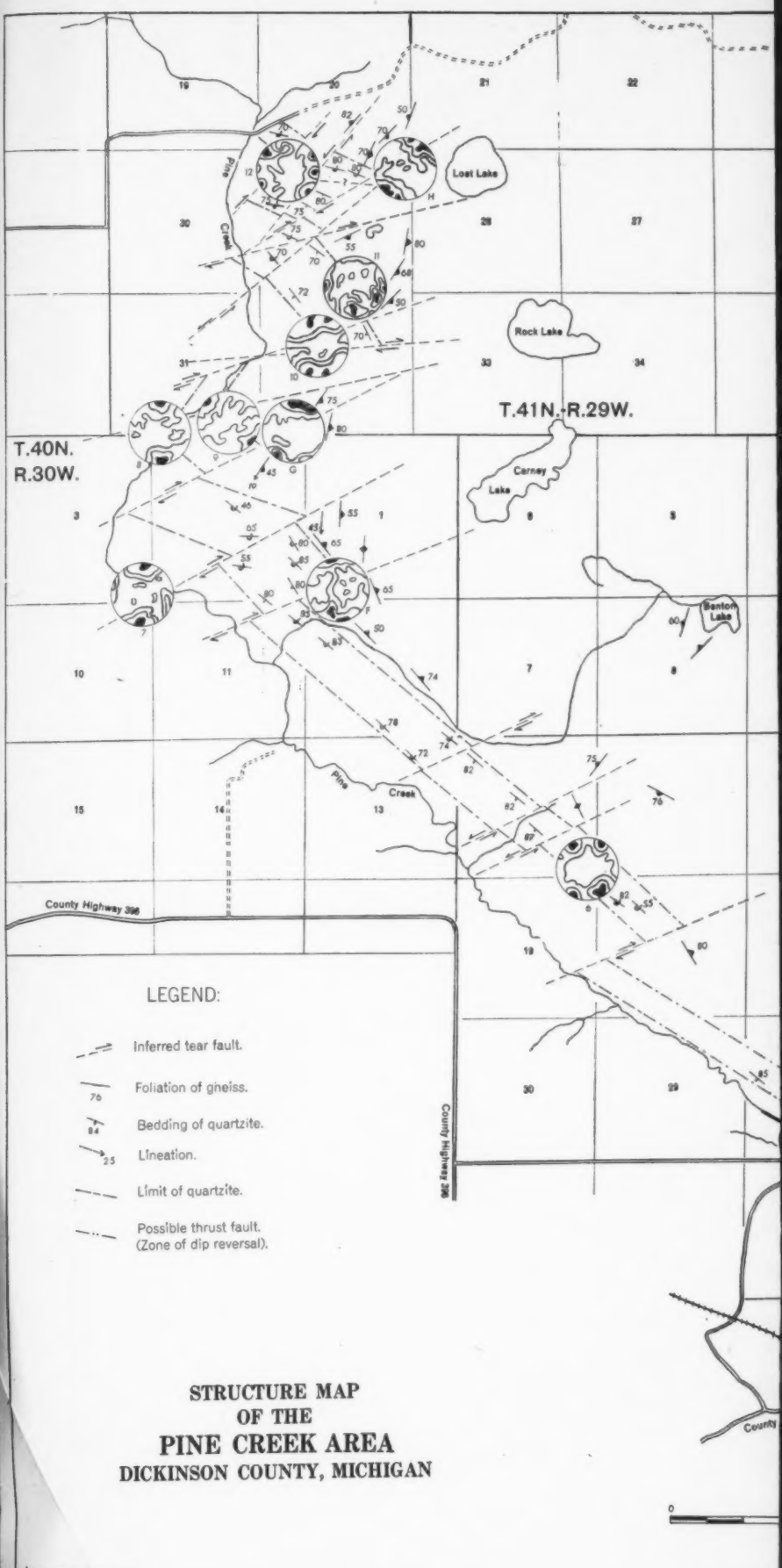
In order to ascertain just how closely the fabrics of the rocks are related to the faulting, two composite diagrams have been made of samples of the faulted southeastern and northwestern areas. The true value of composite diagrams has been pointed out by F. C. Phillips, who states: "These diagrams, it may be noted, summarize facts of observation. Any conclusions which may be derived from them are unaffected by any of the theoretical considerations of structural petrology."[19]

One diagram (Fig. 10) is constructed from 2,600 grains of Samples 1–5 of the quartzite and Samples A–E of the gneiss. The second diagram (Fig. 11) represents 1,950 grains of Samples 7–12 of the quartzite and Samples F–H of the gran-

[18] Pp. 257–62 of ftn. 11.

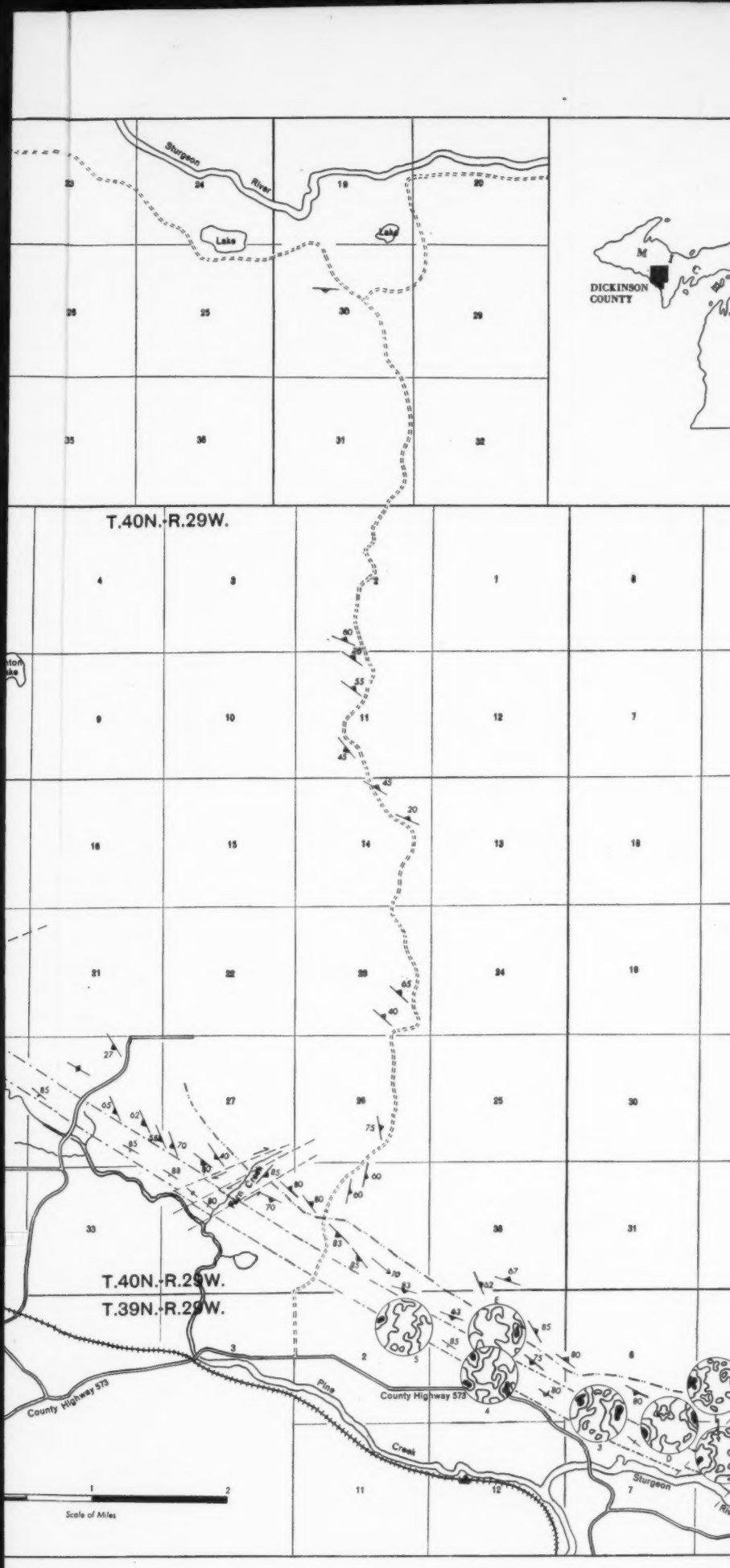
[19] "The Micro-fabric of the Moine Schists," *Geol. Mag.*, Vol. LXXXII (1945), p. 206.

T.4
R.3



Sample

1....300 grains of quartz in quartzite (3 per cent maxima) 8
2....200 grains of quartz in graywacke (4 per cent maximum) 9
3....300 grains of quartz in quartzite (4–5 per cent maxima) 10
4....300 grains of quartz in quartzite (3 per cent maxima) 11
5....300 grains of quartz in quartzite (4 per cent maximum) 12
6....300 grains of quartz in quartzite (4–5 per cent maxima) A
7....250 grains of quartz in quartzite (4–8 per cent maximum) B



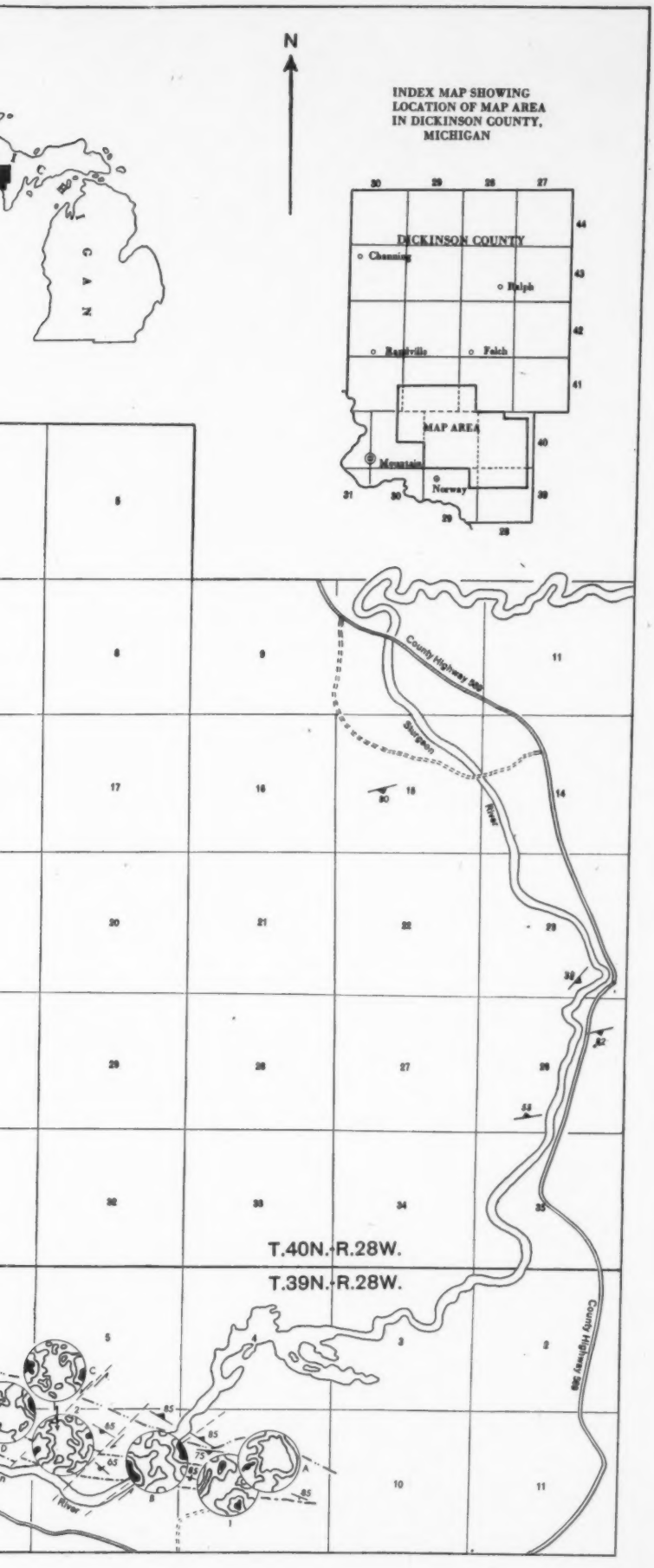
KEY TO FABRIC DIAGRAMS ON STRUCTURE MAP

Sample

8....200 strained grains of quartz in quartzite (4–9 per cent maximum)
9....200 grains of quartz in quartzite (4 per cent maximum)
10....250 grains of quartz in quartzite (4 per cent maxima)
11....250 grains of quartz in green sericitic quartzite (3 per cent maxima)
12....200 grains of quartz in quartzite (4–6 per cent maxima)
A....300 grains of quartz in pink gneiss (4 per cent maximum)
B....200 grains of quartz in pink gneiss (4–6 per cent maximum)

Sample

C....250
D....250
E....200
F....200
G....200
H....200



Sample

C....250 grains of quartz in pink gneiss (4 per cent maximum)
D....250 grains of quartz in gray gneiss (4–5 per cent maximum)
E....200 grains of quartz in pink gneiss (4–5 per cent maximum)
F....200 grains of quartz in gray gneiss (4–6 per cent maxima)
G....200 grains of quartz in brick red granite (4–8 per cent maximum)
H....200 undulatory quartz grains in pink gneiss (4–7 per cent maxima)

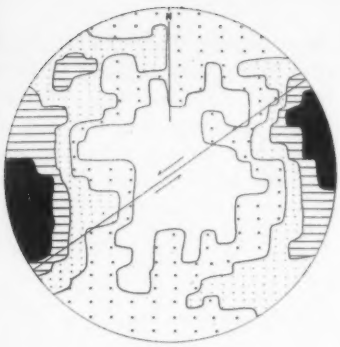


FIG. 10

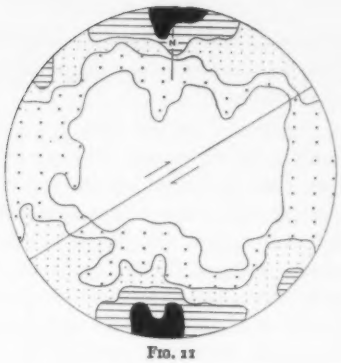


FIG. 11

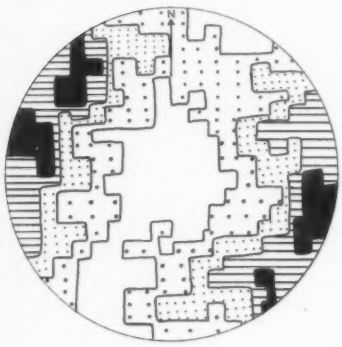


FIG. 12

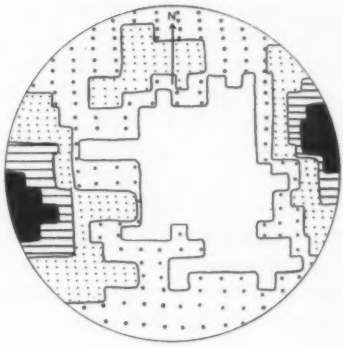


FIG. 13

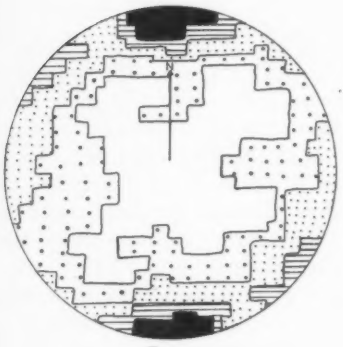


FIG. 14

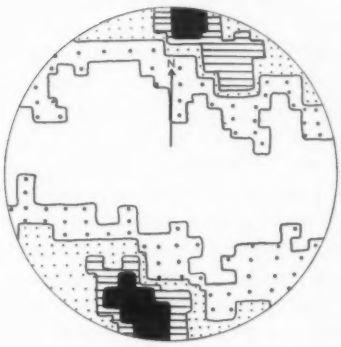


FIG. 15

FIGS. 10–15.—Fig. 10, composite diagram of 2,600 quartz grains of Samples 1–5 and A–E ($\frac{1}{2}$, 1, $1\frac{1}{2}$, 2–$2\frac{1}{2}$ per cent). Southeastern part of Pine Creek area. Fig. 11, composite diagram of 1,950 quartz grains of Samples 7–12 and F–H ($\frac{1}{2}$, 1, 2, 3 per cent). Northwestern part of Pine Creek area. Fig. 12, composite diagram of 1,400 quartz grains of Samples 1–5 of quartzite ($\frac{1}{2}$, 1, $1\frac{1}{2}$, 2 per cent). Southeastern part of Pine Creek area. Fig. 13, composite diagram of 1,200 quartz grains of Samples A–E of gneisses ($\frac{1}{2}$, 1, 2, 3 per cent). Southeastern part of Pine Creek area. Fig. 14, composite diagram of 1,350 quartz grains of Samples 7–12 of quartzite ($\frac{1}{2}$, 1, 2, 3 per cent). Northwestern part of Pine Creek area. Fig. 15, composite diagram of 600 quartz grains of Samples F–H of granite and gneiss ($\frac{1}{2}$, 2, 3, 4 per cent). Northwestern part of Pine Creek area.

ite and gneiss. The average trends of the faults of the areas have been superposed on the diagrams.

The diagram of Figure 10 is a girdle. A peripheral maximum of 2–2½ per cent indicates that a large number of the *c*-axes of the grains are oriented slightly north of east, 25° more to the east than is the average strike of the faults. From the arrows which indicate relative direction of movement of fault blocks, it can be seen that the maximum orientation of the *c*-axes corresponds very well to the direction of elongation of a strain ellipsoid acted upon by a shear couple. The small 1½ per cent submaximum in the northwest quadrant of Figure 10 is possibly due to lateral yielding, through bedding-plane shear, perpendicular to a force from the southwest.

Figure 11 shows that preferred orientation, represented by a 3 per cent peripheral maximum, is north-south. This orientation corresponds to the axis of elongation of a strain ellipsoid acted upon by a couple produced by faulting (Fig. 9); therefore, the maxima are assigned to an *a*-tectonic axis. Fairbairn has cited the development of *a*-maxima in slickensides and states that they "may therefore be considered parallel to the direction of movement."[20] Diagrams of Samples B and H, collected in fault zones, prove the validity of Fairbairn's statement. In the immediate vicinity of the faults the *a*-axis of the strain ellipsoid is tangential to the fault plane. However, it is postulated that, with increasingly greater distances from the fault zone, the *a*-axis of the strain ellipsoid becomes acutely inclined to the strike of the fault. Because most of the rocks studied are located within fault blocks at considerable distances from any fault plane, the maxima of the composite diagrams are developed acutely to the average strike of the faults.

In order to determine whether the fabrics of quartzite are distinctly similar or dissimilar to those of the granitic rocks, separate composite diagrams have been made of the samples of each of these rock types in both the southeastern and the northwestern areas.

The diagrams of the quartzite (Fig. 12) and of the granite (Fig. 13) in the southeastern area are distinctly similar, in that both are girdles with general orientation of quartz axes east-west. A minor difference is that the quartzite diagram has two peripheral maxima, a large one oriented slightly south of east and a smaller one oriented northwest, subparallel to bedding, whereas the diagram of the granitic rocks has but one large peripheral maximum, oriented a few degrees north of east. Preferred orientation subparallel to bedding of the quartzite most likely resulted from bedding-plane shear.

The composite diagrams of samples of the quartzite (Fig. 14) and of samples of granite and gneiss (Fig. 15) of the northwestern Pine Creek area have identical maxima. The only difference between these diagrams is that the maximum of the quartzite diagram is part of a girdle, whereas that of the diagram of the granitic samples is a point maximum.

In summary, it can be stated that the microstructures of rocks of the Pine Creek area are tectonic. Furthermore, all specimens studied have fabrics produced by deformation involving a vertical rather than a horizontal *B*-tectonic axis. Inasmuch as the folding of the Pine Creek area was about a horizontal *B*-axis and the faulting was about a vertical *B*-axis, the rock fabrics are related to the faulting. The general similarity of composite diagrams is adequate proof that

[20] P. 9 of ftn. 12.

the forces involved left the same imprint on both the sedimentary and the granitic rocks.

SUMMARY

Evidences of structure presented in this report are in accord with earlier statements of Bayley and Pettijohn that the complex of granites and gneisses in the Pine Creek area is of Archean age. Trends of foliation of the Archean gneisses are definitely unconformable with the general strike of adjacent upturned Lower Huronian sedimentary rocks. The faults and joints, as well as the microscopic structures of the granitic and sedimentary rocks, are identical.

Also it has been shown that the strata of the Pine Creek area have been deformed at least twice since Archean time, though both deformations probably were separate phases of the Killarney orogeny. During the first phase the Huronian sediments and underlying Archean rocks were folded into a northwest-trending monocline that probably was thrust-faulted along its axial plane. During the second phase of deformation this structure was displaced at its northwestern and southeastern ends by vertical tear faults that trend northeast. The aggregate displacement along faults at the northwestern end has been approximately 3.5 miles. The movement at the southeastern end of the structure was of less magnitude than, and of opposite direction to, that at the northwestern end. The central part of the Pine Creek area was faulted only slightly, but quartzite in this area has been sheared parallel to bedding planes. It is postulated that the faulting and bedding-plane shear have resulted from forces directed from the southwest against the monoclinal structure. These forces were deflected laterally along the central front of the monocline, reinforced by the massive Archean buttress, but were able to displace the monocline at its least resistant ends.

Rock fabrics attest the severity of the fault deformation. The orientation of the *c*-axes of quartz grains in both granitic and sedimentary rocks which crop out in the faulted areas are related to rotational shear about a vertical axis. This shear has been correlated with differential movement of adjacent fault blocks, the components of any one block necessarily being acted upon by a couple.

ACKNOWLEDGMENTS.—The author is grateful to Dr. F. J. Pettijohn for his suggestions and constructive criticisms. He is also appreciative of the able field assistance given by Mr. J. W. Trow and the advice concerning petrofabrics offered by Mr. N. A. Riley.

IGNEOUS METAMORPHISM IN THE OROFINO REGION, IDAHO

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ABSTRACT

A complex series of rocks in the contact-metamorphic zone west of, and adjacent to, the Idaho batholith and underlying the Columbia River basalt is described. The rocks are considered to have been originally chiefly siliceous sediments which have been changed in composition by hydrothermal replacements by solutions from the Idaho batholith to resemble types of quartz diorite and granodiorite. Coarse-grained hornblendite was also formed locally. The over-all textural appearance of these rocks is gneissic, though minor massive and schistose facies occur. Minerals which are considered to have been added to the original siliceous sediments by hydrothermal replacements are biotite, andesine, diopside, uralite, hornblende, scapolite, epidote, zoisite, clinozoisite, allanite, quartz, apatite, sphene, zircon, garnet, magnetite, and sericite.

Ultrabasic rocks in small intrusions, which are described as having occurred somewhat later than the general metamorphism, were altered by autometamorphic, or deuteric, processes.

INTRODUCTION

The purpose of this investigation was to determine the nature and origin of the complex, hornblende-rich, gneissic rocks in the contact metamorphic zone of the Idaho batholith, where it is exposed in the lower Clearwater River Canyon, west of Orofino, Idaho. It was thought that these rocks, because of their high hornblende content, the presence of hornblendite dikelike bodies in them, their megascopic appearance, and their position in relation to recognized gneisses formed by granitization, might be somewhat different in character and origin from previously described granitized sediments. A critical microscopic comparison of these rocks with the recognized gneisses formed by granitization in other parts of the region was made, in order to obtain a better insight into their origin. Some small ultra-basic intrusives were studied in less detail to determine their relationship, if any, to the problem.

Field investigation was made during the summers of 1941 and 1942, and the microscopic examination of rock sections was made during the academic years 1941–42 and 1942–43.

The area which is included in this study lies in the western part of Clearwater County and the northeastern part of Nez Perce County, Idaho, and centers about the town of Orofino. That portion of the area which was studied in detail lies along three main streams—the Clearwater River, from about 5 miles south of Orofino to 5 miles west of Peck Station; the North Fork, from Ahsahka to about 5 miles east of Elk Creek; and Orofino Creek, 20 miles eastward from Orofino. The location and boundaries are shown in Figure 1.

GENERAL GEOLOGIC SETTING

The Orofino region is underlain by a thick, complex series of sedimentary rocks, which were metamorphosed by the intrusion of the Idaho batholith during late Mesozoic time.[1] The metamorphic zone is broad and extends at least 30 miles west of the main outcrops of recognized batholithic rocks, indicating that the batholith broadens below the surface and probably does not lie at great depth.

[1] C. P. Ross, "Mesozoic and Tertiary Granitic Rocks in Idaho," *Jour. Geol.*, Vol. XXXVI (1928), pp. 673–93; A. L. Anderson, "The Geology and Mineral Resources of the Region about Orofino, Idaho," *Idaho Bur. Mines and Geol., Pamphlet 34* (1930), p. 18.

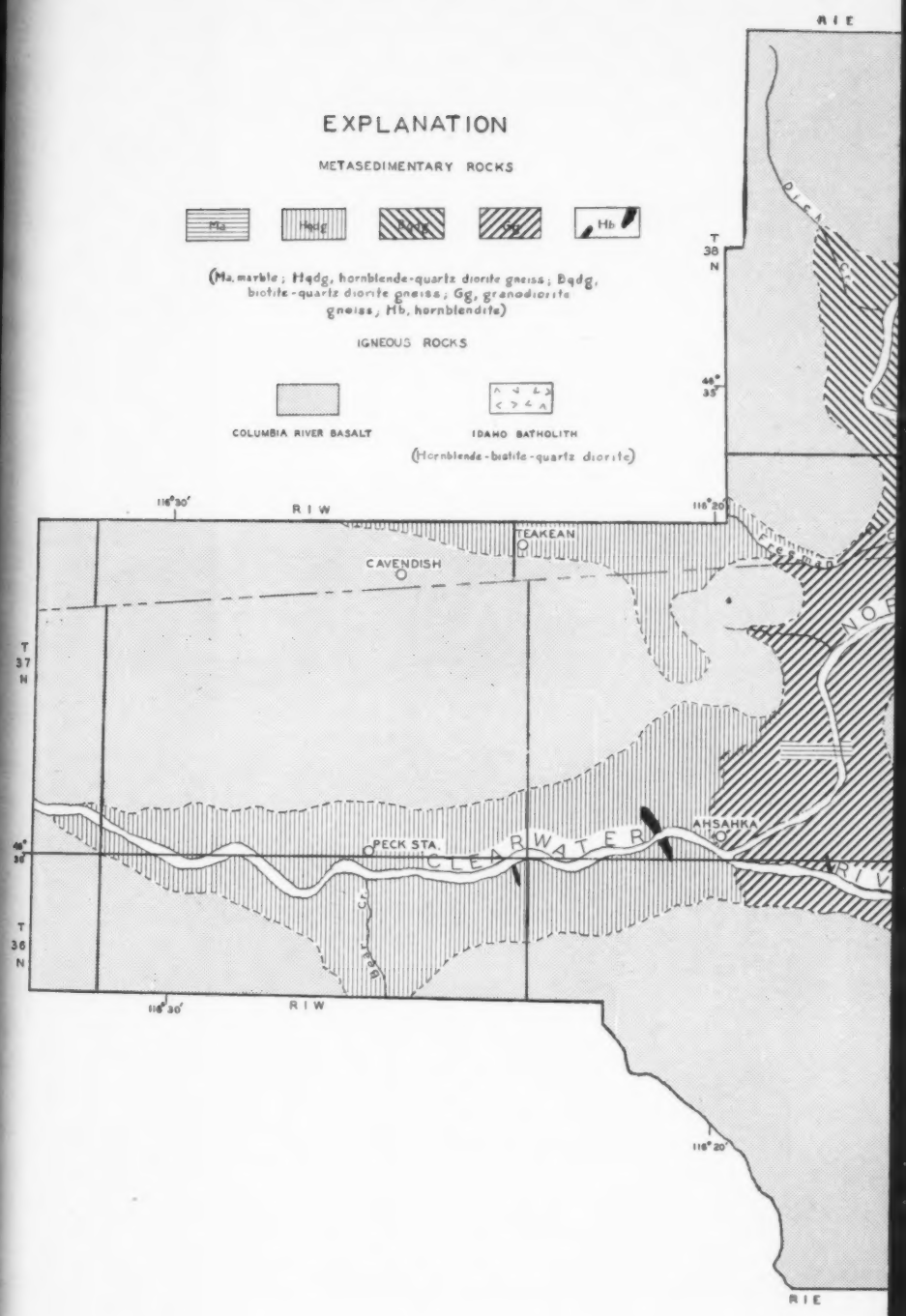
EXPLANATION
METASEDIMENTARY ROCKS
Ma
Hqdg
Bqdg
Gg
Hb
(Ma, marble; Hqdg, hornblende-quartz diorite gneiss; Bqdg, biotite-quartz diorite gneiss; Gg, granodiorite gneiss; Hb, hornblendite)
IGNEOUS ROCKS
COLUMBIA RIVER BASALT
IDAHO BATHOLITH
(Hornblende-biotite-quartz diorite)
R 1 E
T 38 N
46° 35'
116° 30'
R 1 W
116° 20'
TEAKEAN
CAVENDISH
T 37 N
PECK STA.
CLEARWATER R.
AHSAHKA
T 36 N
46° 30'
R 1 E

FIG.

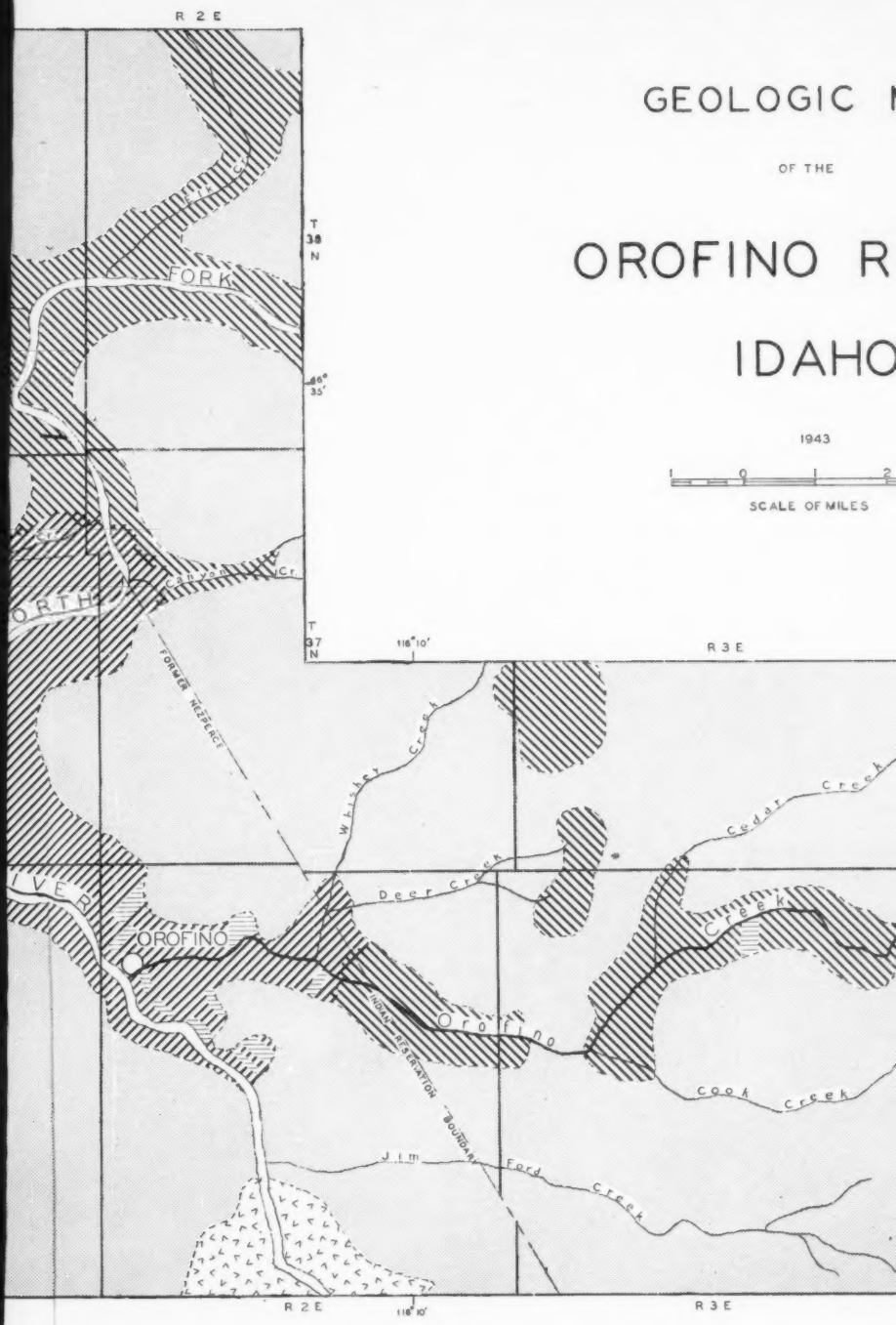

FIG. 1.—Geologic map of the Orofino Region, northern Idaho

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REGION,

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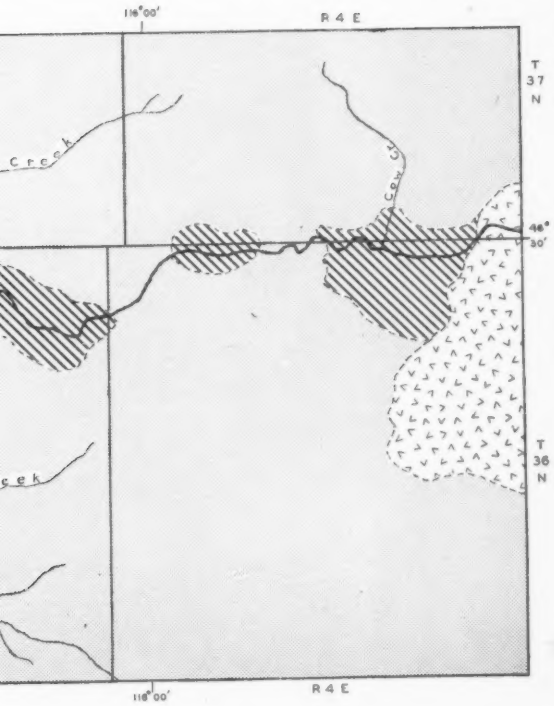


The metasedimentary rocks were given the name "Orofino series" and were assigned to the lower Belt series of pre-Cambrian age by A. L. Anderson in 1930,[2] but he now believes that they more probably are Paleozoic or early Mesozoic in age.[3] After the intrusion of the batholith, erosion cut deeply into the metasediments and exposed a large area of the batholith.

During the Miocene epoch, the Columbia River basalt flowed over this irregular erosion surface, covering the metasediments and the margins of the batholith to a depth of 2,000 feet or more. Since that time, basalt has covered the greater part of the area, and the older rocks are exposed only where the streams have cut deep canyons through the basalt and in a few isolated areas where the old erosion surface stood exceptionally high.

The rocks primarily dealt with in this report are the metasediments which crop out along the Clearwater River, the North Fork, and Orofino Creek. They are mainly gneisses, which vary in composition from hornblende-biotite–quartz diorite to granodiorite. A peculiar phase of the hornblende-rich rocks, known as "hornblendite," occurs as dikelike bodies in the gneiss; and coarsely crystalline, pure to silicated marble layers are intercalated with the gneisses. Small bodies of dunite also occur locally in the gneiss.

Detailed structure of the metasedimentary rocks cannot easily be determined because the extent of the exposures is limited by the overlying basalt and talus from that basalt. However, exposures of small extent are plentiful, and, although the rocks have been altered and appear granitic in texture, the original bedding has not been entirely obliterated. Consequently, fairly accurate dip-and-strike measurements can be obtained. The strike ranges in general from east-west to N. 45° W. but varies from due west to east of north. The dip ranges from 15° to 85° to the north. Almost all the rocks have been intricately broken, but no apparent regular fracture pattern could be worked out.

Abrupt changes along the strike from marble to gneiss suggest that transverse faulting may have occurred.

Petrography

Introduction

The rocks of the area are divided into two main groups, igneous and metasedimentary. The igneous rocks are quartz diorite in the Idaho batholith and dunite in small intrusions into the metasediments. The metasedimentary rocks, except marble, have the composition and the textural appearance of igneous rocks and have been so called. However, the rocks occur in distinct, thick to thin, parallel layers which greatly resemble bedding and which, outside this area,[4] grade into almost pure quartzites, on the one hand, and igneous rocks, on the other. Moreover, microscopic examination shows early quartz occurring in granular, more or less equidimensional, interlocking grains as in quartzites and siliceous shales, which indicates that the present composition and texture of the rock are the result of hydrothermal metamorphism of siliceous sediments. Because the metasedimentary rocks, except the marble, have the composition of igneous rocks, the author has given them igneous rock names. These are, in the order in which they will be described, hornblende–quartz diorite gneiss, hornblendite, biotite–quartz diorite gneiss,

[2] P. 9 of ftn. 1.

[3] Personal communication.

[4] A. L. Anderson, personal communication.

granodiorite, and marble. All these rocks, except the marble, are gradational one into another and have been separated on the basis of evidence furnished by microscopic study of thin sections.

IGNEOUS ROCKS

IDAHO BATHOLITH—QUARTZ DIORITE

A large mass of granitoid rock, commonly known as the "Idaho batholith," lies to the east and south of the Orofino region. This batholith has been described[5] as being composed of "two main facies which differ from one another in age, structure, and composition. One facies is represented in a broad marginal and roof zone of more calcic rock . . . .; the other is a less calcic inner zone. . . . ."

Granitic rocks, believed by the writer to be part of this marginal facies of the batholithic mass and to be genetically related to the metasedimentary rocks of the Orofino region, occur at the southern and eastern edges of the area covered in this report (Fig. 1). The rock at the eastern edge of the area is biotite-bearing quartz diorite and is very much like those of the marginal facies referred to above. The rock to the south is similar also but differs in that it contains a rather high percentage of hornblende and a low percentage of quartz.

Since these rocks have been described in detail by Anderson, a brief summary of his petrographic description is given here as follows:

> The marginal facies is mostly quartz diorite. The rock is moderately dark gray, and the granularity is medium to coarse. The essential minerals are sodic andesine (45–60 per cent), quartz (15–30 per cent), and biotite and hornblende (10–15 per cent). Microcline is usually present in very small amounts but locally forms 15–30 per cent of the rock. Sphene, epidote, apatite, magnetite, zircon, allanite, zoisite, chlorite, muscovite, and sericite also were present.

[5] A. L. Anderson, "Endomorphism of the Idaho Batholith," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 1102–7.

ULTRA-BASIC ROCK—DUNITE

Field relationships.—Small, irregular bodies of dunite occur in the metasedimentary gneisses. The known areas of outcrop are so small that it is impossible to show them accurately on the accompanying geologic map (Fig. 1). The three occurrences studied are about 3 miles southeast of Teakean, where exposures occur along an area about 10 feet wide an 100 feet long; about 2 miles northwest of Orofino, on the northeast side of the Clearwater River, where a body 10–15 feet wide and 20–25 feet long is exposed; and about 14 miles east of Orofino in a railroad cut, where a dike 10–15 feet wide crops out for ½ mile, disappearing at each end under basalt and taulus. The dunite cuts across the gneisses at varying angles. Near Teakean, where the best exposures were found, the gneisses strike N. 80° W. and dip 75° northeast, whereas the dunite strikes N. 60° W.

Megascopic description.—The dunite is black to green if relatively fresh but becomes greenish-gray to light gray on alteration. It is massive and fine- to medium-grained. Partially altered rock is flecked with areas of fibrous and foliated minerals, which increase in size with progressive alteration until the original structure of the rock is obliterated. The elongated crystals range from fine fibers less than ¼ inch in length to coarse columnar crystals 20 inches long. The foliated material is matted together in aggregates, and therefore the size of individual crystals cannot be determined. Recognizable mineral constituents are olivine, amphibole, and talc, which occur in variable amounts, depending on the stage of alteration attained.

Microscopic description.—The granularity of the fresh rock is generally fine to medium, and the grains are equidimensional, giving the rock its massive structure. The altered rock is fine- to coarse-grained, and the proportion of fibrous to more or less equidimensional minerals varies considerably. Olivine, the chief mineral constituent of this rock, ranges from 100 to less than 5 per cent of the minerals present. The other mineral constituents are amphibole, chlorite, talc, serpentine, quartz, apatite, magnetite, and carbonate.

Olivine occurs in fractured grains ranging from less than 0.25 up to 5 mm. in diameter. The olivine is locally altered to a deep-brown mineral, with high relief and aggregate structure, which was identified as iddingsite. Amphibole, talc, and serpentine penetrate the olivine and isolate grains and parts of grains, accentuating the fineness of the granularity of the rock.

Iddingsite occurs in the dunite east of Orofino. It is dark-brown in color, and some grains show faint pleochroism; it has high relief and displays an aggregate structure. This mineral is an alteration of the olivine and takes the form of the olivine grains when alteration is complete. The change begins along cracks, where the iddingsite has a somewhat fibrous habit. Elsewhere it is fibrolamellar. The writer is not certain of the paragenetic position of the iddingsite. It replaces the olivine and seems to be penetrated by the other minerals as stated. However, it is recognized that the iddingsite may mold itself around the other minerals, as it replaces the olivine for which it seems to have a specific affinity. The water content of iddingsite favors the latter interpretation.

Amphibole occurs as coarse prismatic crystals and as fine prismatic to fibrous crystals, which grow in matted aggregates. The optical properties conform to those of both tremolite and anthophyllite. The grains are colorless and show moderate birefringence. The extinction is parallel or inclined as much as 18°. The crystals are length-slow and display both positive and negative interference figures with large axial angles. The fine material conforms more closely to anthophyllite, which can have inclined extinction in sections that cut all three crystallographic axes. Age relationships between the two are indistinct, but the fibrous material seems to cut across the coarser grains, which conform more closely in optical properties to tremolite. Both seem to pierce olivine and iddingsite, and both appear to be replaced by talc and serpentine, which occur as thin finger-like projections along cleavage directions and fractures in the amphibole.

Colorless to pale-green chlorite occurs in irregular to sheaflike masses. It seems to grow against, and to send minute stringers into, amphibole. Opposite relationships of chlorite to talc are indicated. However, because of the habits of chlorite and talc and because they are probably near the same age, their paragenetic relationships are obscure. Furthermore, the growth of serpentine between chlorite and talc tends to destroy the boundaries between these two minerals.

Talc occurs as very fine aggregates and as flakes up to 1.5 mm. across. It penetrates olivine, amphibole, and possibly chlorite. Serpentine has the appearance of eating into the larger flakes of talc, in places isolating parts which remain in optical continuity with one another. It also sends fine tongues into some smaller flakes of talc and entirely engulfs others.

Long, fibrolamellar crystals of serpentine form in cracks in the olivine and in places surround and isolate parts of oli-

vine grains. In this process fine projections of serpentine penetrate into the olivine, perpendicular to these cracks. Amphibole, chlorite, and talc also are attacked along cleavage directions and elsewhere by serpentine, which penetrates grains and isolates parts of grains.

Quartz, apatite, magnetite, and carbonate form a minor part of the rock, and their detailed paragenesis has not been completely established. Quartz is found as veinlets in cracks or fillings between the other minerals. Magnetite occurs as fine grains or as dust in masses of chlorite, the formation of dust accompanying the development of the chlorite. Apatite and carbonate are rare.

METASEDIMENTARY ROCKS

HORNBLENDE–QUARTZ DIORITE GNEISS

Field relationships.—Hornblende–quartz diorite gneiss crops out along the Clearwater River almost continuously from near Ahsahka to about 4 miles west of Peck Station, where it disappears beneath basalt (Fig. 1). Quartz veins and several hornblendite bodies cut across this gneiss. Pegmatite dikes of composition similar to the gneiss are abundant. The boundaries of the hornblende–quartz diorite gneiss are gradational with those of the granodiorite gneiss, the biotite-quartz gneiss, and the quartz diorite of the Idaho batholith.

Megascopic description.—This rock is medium to dark gray. The granularity is medium to coarse, but fine-grained schistose phases occur locally. Rather distinct gneissic structure is noticeable in outcrops but is not so apparent in hand specimens. Minerals which can be identified megascopically are feldspar, hornblende, biotite, quartz, garnet, and pyrite. Feldspar is generally less abundant in the schistose phases than in the gneiss proper and decreases in amount in the gneiss near the west end of the area. Hornblende increases in amount near the hornblendite dikes and in the rocks west of Peck Station.

Microscopic description.—The hornblende–quartz diorite gneiss is inequigranular-hypidiomorphic. Most of the grains are embayed and penetrated by adjoining minerals, and some large grains tend to engulf smaller ones. Gneissic structure is pronounced to weak, depending on the abundance of feldspar and hornblende porphyroblasts, which tend to grow across and destroy this structure.

Quartz, biotite, plagioclase, and hornblende constitute the major part of this rock. Allanite, epidote, zoisite, clinozoisite, chlorite, sericite, apatite, sphene, zircon, garnet, pyrite, and magnetite occur in variable, but generally minor, amounts.

Quartz occurs in two different habits. The first of these is in small, irregular, more or less equidimensional, interlocking grains, much like that in quartzites and siliceous shales, and is interpreted as remnants of siliceous sediments (Pl. I, figs. 2 and 3).

Biotite also shows dual relationships to the other minerals. Irregular flakes of earlier biotite show rather good alignment. These flakes may lie across several grains, occur between grains, or pierce grains of the early quartz. Because of the relationship of biotite to early quartz and because of field evidence to be discussed under "Petrogenesis," it is considered to be a later mineral than this quartz. The alignment of the biotite is the main cause of the gneissic structure of the rock.

Small, blocky to irregular grains of twinned plagioclase with the composi-

tion of soda andesine near $Ab_{70}An_{30}$ occur in the quartz mosaics and differ in habit and in relationships from andesine of a later stage. Prongs of the irregular grains grow between early quartz and early biotite and penetrate and encircle grains of these minerals, which may become entirely engulfed in the andesine (Pl. I, figs. 3 and 5). Many of the boundaries between this early andesine and early biotite are sharp, but others are ragged, where minute seams of andesine lie along cleavage directions in the biotite. These spatial relationships of early andesine to early biotite and early quartz indicate that it is a later mineral than either of them. Early andesine occurs as inclusions in the second form of andesine, but more details of this relationship will be given below.

Hornblende and biotite constitute 20–40 per cent of the rock in the approximate proportion of 4 to 1. The hornblende shows distinct pleochroism—greenish-gray when the cleavage traces of longitudinal sections are parallel to the vibration direction of the lower nicol and yellowish-green when these traces are perpendicular to this position. Anhedral to subhedral grains as large as 7 mm. in length lie parallel to and across the gneissic structure. Irregular tongues and protuberances of hornblende lie between and penetrate grains of early quartz, early biotite, and early andesine. These three minerals may lie in contact with hornblende or occur in the hornblende entirely isolated from other like grains, giving evidence of the growth of hornblende around them (Pl. I, fig. 1). Late andesine, late biotite, minerals of the epidote group, chlorite, late quartz, apatite, sphene, zircon, and magnetite encroach upon and lie in cracks and cleavages in the hornblende.

Porphyroblastic andesine, also about $Ab_{70}An_{30}$, is more abundant than the fine-grained early andesine described above. Together they constitute 40–60 per cent of the rock. This andesine is in grains as large as 6.5 mm. in length. It shows both albite twinning and pericline twinning. The crystals are subhedral in outline and penetrate irregularly into and between the early quartz, the early andesine, and the early biotite, resulting in all stages of engulfing composites or individual grains of these three minerals. In early stages of growth the late andesine forms lacy, skeletal crystals, the outlines of which are barely discernible in the quartzite (Pl. I, fig. 2). More mature crystals may hold a few isolated grains of early quartz, early biotite, early andesine, and/or hornblende, having replaced all the original intervening material. Porphyroblasts of late andesine, which have very ragged borders because of irregular growth into and between surrounding finer-grained minerals, give the appearance of having been partially crushed; but the diversity of minerals in these fine-grained borders, the absence of cataclastic phenomena, and the presence of replacement relations show that they are not fragments from larger grains (Pl. I, fig. 3). Late andesine also grows into hornblende crystals both across and along cleavages and isolates corroded remnants of hornblende, which remain in optical continuity with the parent-crystal (Pl. I, fig. 4). All these relationships tend to show the replacement character of the late andesine and indicate that it is later than early quartz, early biotite, early andesine, and hornblende. Alteration areas border and occur in many of the andesine grains (Pl. I, fig. 6 and Pl. II, fig. 5). They broaden near and extend along cracks and cleavages, giving the andesine a mottled appearance. Patches of similar material

form, apparently independently of the border areas. However, the border areas and patches are all in optical continuity throughout one andesine crystal, and they all have smaller extinction angles than does the unaltered andesine. Minerals of the epidote group commonly lie inside these altered areas in the andesine. Apparently, cooling solutions penetrated between grains of, and along cracks and cleavages in, this late andesine, changing the andesine to a more sodic plagioclase and using the elements released in the process to form the minerals of the epidote group.

A second-generation biotite differs from the first in habit and penetrates and corrodes the minerals which precede it. Small, ill-formed bodies develop thin, translucent smears and colorations in hornblende; and thin, delicate seams form along cleavage directions and fractures in hornblende and andesine. Larger flakes send fingers between the previously formed minerals. Fine, needle-like projections on the early biotite might be a growth of late biotite, but sufficient evidence for proof of this could not be found.

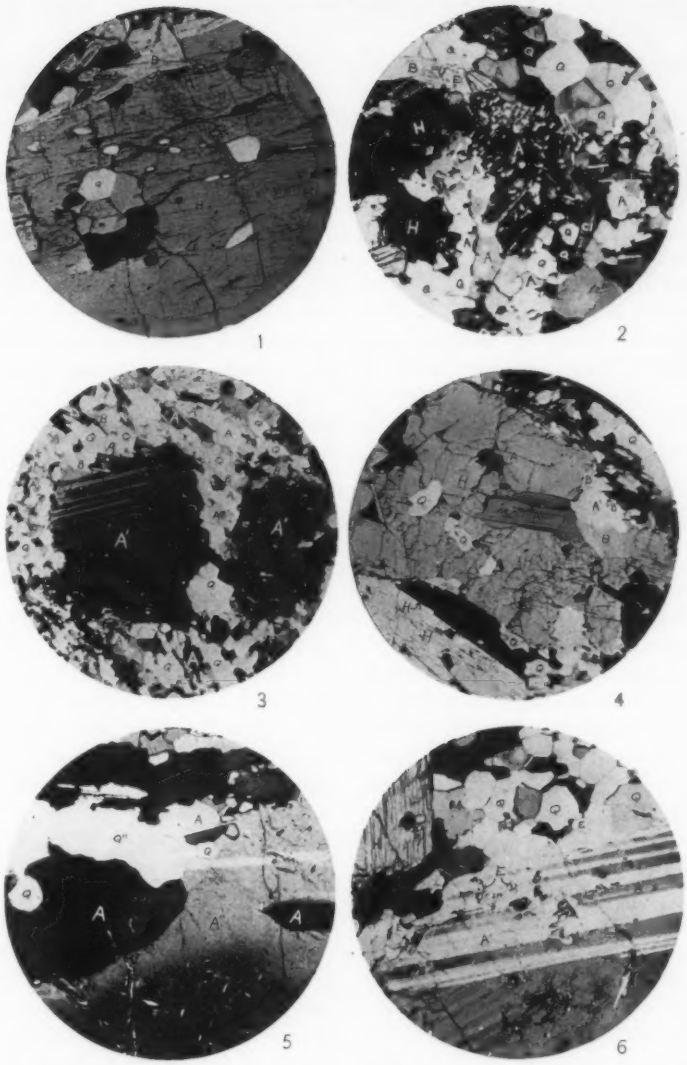
Allanite, epidote, zoisite, and clinozoisite occur in variable amounts, locally forming 10 per cent or more of the rock. All four minerals grow in andesine grains, commonly along cleavage directions and cracks or haphazardly from the sides, and generally lie in or near the alteration areas which were described above (Pl. I, fig. 6, and Pl. II, fig. 5). These minerals also form blocky and irregular grains in hornblende and biotite, apparently growing across the crystals indiscriminately. Allanite commonly occurs as cores around which epidote has grown, but it also occurs by itself. The association of the minerals of the epidote group with the reaction areas in the andesine and the irregular growth in hornblende and biotite crystals suggest that they have formed from the reaction of cooling solutions on andesine and possibly on hornblende and biotite.

Chlorite generally occurs in straight sheaflike, or irregular masses in hornblende and biotite, but it is also present in andesine, in which included biotite has been altered to chlorite. It is scarce, and its paragenetic relationships to minerals other than hornblende and biotite are obscure.

Late quartz differs from the early granular quartz in its habit and spatial relationships to the other minerals. It commonly forms intergranular veinlets and irregular bodies and sends tongues into fractures in adjacent minerals. Large lobate protuberances also extend from the veinlets and irregular bodies and indent andesine and hornblende crystals (Pl. I, fig. 5). Veinlets of late quartz grow irregularly along cleavage planes in hornblende, giving it a motheaten appearance. Early and late quartz together form 5–20 per cent of the rock.

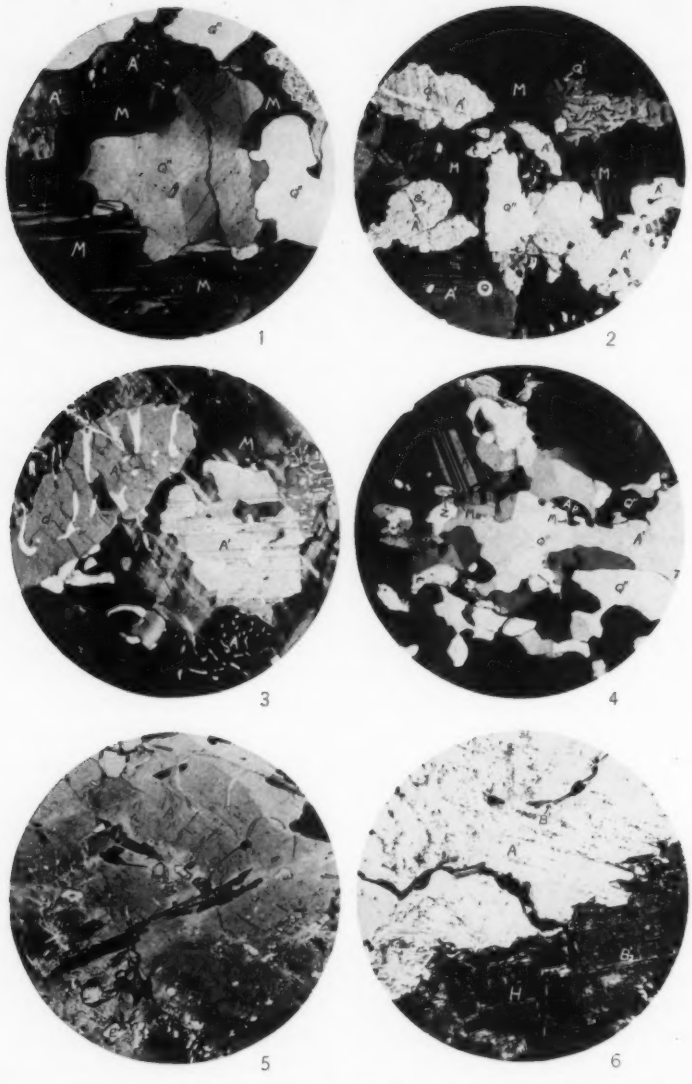
Apatite, sphene, and zircon occur in variable, but minor, amounts. All are closely associated with late quartz and commonly occur as small grains along cracks in the quartz veinlets. In addition, they occur in intergranular spaces and in cracks in other minerals. Apatite also forms blunt, elongate grains, which grow haphazardly into andesine, hornblende, and biotite. Sphene forms large to small irregular grains growing through and around hornblende and other minerals. Zircon occurs as small to minute, subhedral to round, more or less elongate crystals, which are confined chiefly to cracks. There is no direct evidence to show the age relationship between apatite and sphene, but zircon grows along cracks in both these minerals.

PLATE I



Photomicrographs of the hornblende-quartz diorite gneiss. *1*, A hornblende porphyroblast (H) with remnants of early quartz (Q) and early biotite (B) as inclusions. Epidote (E) also is present. Mag. ×50. *2*, A porphyroblast of late andesine (A′) in early stage of growth replacing early quartz (Q) and early andesine (A). Hornblende (H), early biotite (B), epidote (E), and clinozoisite (Cl) also are present. Mag. ×50. *3*, Late andesine porphyroblasts (A′) growing in a mosaic of early quartz (Q), early andesine (A), and early biotite (B). The apparent granulation is a pseudoclastic effect resulting from the replacement of the mosaics by andesine. Mag. ×20. *4*, Late andesine (A′) and late biotite (B′) replacing hornblende (H) which had previously surrounded early quartz (Q), early andesine (A), and early biotite (B). Mag. ×20. *5*, Late andesine (A′) includes a remnant of early andesine (A) in optical continuity with the parent grain, which holds early quartz (Q). Late quartz (Q″) embays both andesines. Mag. ×50. *6*, Epidote (E) surrounded by a reaction area which cuts twinning lamellae in late andesine (A′). Early quartz (Q), hornblende (H), and late biotite (B′) are also shown. Mag. ×50.

PLATE II



Photomicrographs of the granodiorite gneiss and (*5*) of hornblende-quartz diorite gneiss and (*6*) of hornblendite. *1*, Late quartz (Q″) embaying and surrounding late andesine (A′) and microcline (M). Myrmekitic quartz lies in the late andesine in the upper part of the illustration. Mag. ×50. *2*, Microcline (M) isolating late andesine (A′) in which myrmekitic quartz (Q) is growing. Early quartz (Q) is engulfed by late andesine. Late vein quartz (Q″) is also present. Mag. ×50. *3*, Myrmekitic quartz (Q′) in late andesine (A′) which is surrounded and penetrated by microcline (M), which also surrounds but does not replace the myrmekitic quartz as is illustrated in the extreme left part of the figure. Sericite (Se) pierces microcline, late andesine, and myrmekitic quartz. Mag. ×50 *4*, Magnetite (M) surrounds zircon (Z) and apatite (Ap) and penetrates late andesine (A′). Late quartz (Q″) also replaces late andesine. Mag. ×50. *5*, Late andesine (A′) partially altered to epidote (E) and clinozoisite (Cl), which are surrounded by reaction areas (light) in the andesine. Late biotite (B′) also is present. Mag. ×50. *6*, A ragged replacement boundary between late andesine (A′) and hornblende (H) in the hornblendite. Late biotite (B′) is present in both the hornblende and the andesine. The irregular dark line diagonally across the center of the figure is a crack. Mag. ×20.

Subhedral crystals of garnet and pyrite frequently occur together along fairly definite zones parallel to the gneissic structure and where quartz veinlets are common.

Magnetite is in grains of variable size and shape. It grows as blunt to long narrow tongues in interstitial cracks between the other minerals, seemingly widening these cracks during the process. The boundaries of magnetite with almost all the other minerals are generally irregular to ragged. The magnetite also forms narrow, fringelike borders around apatite, sphene, and zircon, somewhat destroying their crystal outlines. Long, irregular tongues and prongs of magnetite grow into and across andesine, hornblende, biotite, and the other minerals, showing the replacement character of the magnetite.

HORNBLENDITE

Field relationships.—Four occurrences of rock, described as hornblendite by A. L. Anderson,[6] are shown on the map (Fig. 1). Hornblendite occurs as dikelike bodies cutting the metasedimentary rocks at various angles to the original bedding. Three of the bodies are less than 20 feet wide, and the other is about 200 feet wide. Numerous other smaller bodies occur in the area, but they are so small that they are not described specifically. All these bodies occur in the hornblende–quartz diorite gneiss. The lengths of the dikes could not be determined because they disappear beneath the basalt talus within short distances. The largest dike crops out on both sides of the Clearwater River about 1 mile west of Ahsahka. The contacts between the dikes and the adjacent gneiss are gradational, and generally the hornblende content of the gneiss increases toward the dikes. Apophyses of hornblendite extend into the gneiss for distances of 3–15 feet from the main bodies.

Megascopic description.—The hornblendite is dark gray to black, very coarse grained, and massive. Ordinarily, the minerals, especially the hornblende, decrease in size toward the margin of the dikes, where they grade into gneisses. The crystals in the small dikes range from less than 1 to 6 inches in length, and in the large dike from 1 to 14 inches in length. The grains of other minerals are not so large as those of the hornblende, but some of the plagioclase crystals are 10–15 mm. across. Stringers and lenses of white feldspar fill cracks in the hornblende. Biotite flakes in discontinuous cross-fractures impart a coarse schillerlike structure to the hornblende.

Microscopic description.—The hornblendite is an extremely coarse-grained, massive rock with hypidiomorphic-granular texture. All the minerals in the hornblendite also occur in the hornblende–quartz diorite gneiss, but the reverse is not true. The relationships of the minerals are the same, but the size of the crystals and the proportions of minerals are different. The minerals of the hornblendite proper are hornblende, biotite, plagioclase, allanite, epidote, chlorite, sericite, quartz, apatite, and magnetite.

Hornblende constitutes 75–100 per cent of the rock. The crystals are very large and have good cleavage. The pleochroism is like that of the hornblende in the adjacent gneiss—greenish-gray when the cleavage traces in longitudinal sections are parallel to the vibration direction of the lower nicol and yellowish-green when these traces are perpendicular to this position. Irregular cracks in this mineral are filled with plagioclase and quartz, probably analogous to the late andesine

[6] "An Occurrence of Giant Hornblendite," *Jour. Geol.*, Vol. XLI (1933), pp. 89–98.

and the late quartz of the hornblende–quartz diorite gneiss, as explained in the following paragraphs. Regular, interrupted cross-fractures at about right angles to the cleavage traces are filled with biotite, probably late biotite, as explained below. Near the margins of the dikelike bodies, hornblende includes remnants of early quartz and early andesine, as described in the hornblende–quartz diorite gneiss. The character of the hornblende, the gradual decrease in hornblende content in the gneiss away from the hornblendite bodies, and the replacement relationships with minerals of the gneiss in the gradational margins of the bodies indicate that the hornblendite represents a localized, special phase of hornblendization in the general process of granitization.

The plagioclase occurs as large crystals, which show very coarse but poorly defined twinning. This plagioclase is soda andesine of about the same composition as the late andesine in the adjacent gneisses. It occurs as veinlets and vug fillings in the hornblende. The boundaries between andesine and hornblende are irregular, because the andesine has made embayments into hornblende and has sent thin seams along the cleavage directions (Pl. II, fig. 6). Isolated grains of hornblende in optical continuity with near-by large crystals are found in the andesine. Near the margins of the dikes, this andesine penetrates and engulfs early quartz, early biotite, and early andesine, as does the late andesine in the gneiss. These relationships tend to show that the andesine in the hornblendite is of the same stage as the late andesine in the hornblende–quartz diorite gneiss.

The habit of the biotite in the hornblendite is similar to that of the second-generation biotite in the hornblende–quartz diorite gneiss. It forms smears, small ill-defined bodies, and large flakes in the hornblende. Other small flakes of biotite fill regular, but interrupted, cross-fractures in the hornblende. In addition, thin blades of biotite occur along cleavage directions and fractures in the andesine (Pl. II, fig. 6). Allanite and epidote appear to have about the same relationships to other minerals that they have in the adjacent gneiss. They grow along fractures and penetrate and lie in andesine, hornblende, and biotite. No evidence to show that epidote is associated with reaction areas in the andesine of this rock was observed. However, the abundance of sericite in the andesine where epidote occurs may have tended to obliterate this evidence. Other relationships are obscure because allanite and epidote are not abundant.

Chlorite grows in irregular and sheaflike aggregates and is rather abundant, compared to its occurrence in the adjacent gneiss. Definite growth relationships between chlorite and minerals other than hornblende and biotite could not be determined. It grows into the hornblende and biotite along cleavage traces and at the margins of those minerals.

Late quartz occurs much as does similar quartz in the adjacent gneiss but in smaller quantities. It forms veinlets and lobate grains, which penetrate hornblende, andesine, and biotite.

Apatite grows between hornblende and andesine, projecting into both, and occurs in cracks in the andesine. Its association with late quartz is not definite.

Magnetite occurs as rather large irregular grains, which send tonguelike projections along intergranular cracks and pierce adjacent minerals.

BIOTITE–QUARTZ DIORITE GNEISS

Field relationships.—The main outcrop areas of biotite–quartz diorite gneiss are along Orofino Creek, east of the mouth of Whiskey Creek, and along

the North Fork, north of the mouth of Freeman Creek. Two small, isolated areas occur between Orofino and Whiskey Creek (Fig. 1). Quartz veins and pegmatite dikes of composition similar to the gneiss are abundant. The boundaries of the biotite–quartz diorite gneiss are gradational into the granodiorite gneiss and into the quartz diorite of the Idaho batholith. Marble is locally intercalated between layers of the gneiss.

Megascopic description.—This rock is light to moderately gray in color, not so dark as the hornblende–quartz diorite gneiss. The granularity is generally medium to coarse. Gneissic structure is distinct in most outcrops, but some hand specimens appear to be massive. Minerals which can be identified in hand specimens are feldspar, biotite, quartz, hornblende, and garnet. Feldspar, biotite, quartz, and garnet are more abundant than in the hornblende–quartz diorite gneiss.

Microscopic description.—In thin section the biotite–quartz diorite gneiss resembles the hornblende–quartz diorite gneiss in texture, structure, mineral composition, and paragenetic relations. The minerals present are quartz, biotite, andesine, hornblende, allanite, epidote, zoisite, clinozoisite, chlorite, apatite, sphene, zircon, garnet, pyrite, magnetite, and sericite. The chief difference between the two rocks lies in the proportion of minerals. The biotite–quartz diorite gneiss contains a lower percentage of mafic minerals in general, a higher percentage of biotite, and slightly higher percentages of andesine and quartz. These differences are shown in Table 1.

TABLE 1

COMPARATIVE AMOUNTS OF MAJOR MINERALS IN THE GNEISSES AND THE HORNBLENDITE

Name of Rock	Andesine (Per Cent)*	Microcline (Per Cent)*	Quartz (Per Cent)*	Mafic Minerals (Per Cent)*	Biotite-Hornblende Ratios*
Hornblende–quartz diorite gneiss	40–60	.....	5–20	20–40	1:4
Hornblendite	0–25	.....	Less than 5	75–100	1:90 or less
Biotite–quartz diorite gneiss	45–65	.....	10–25	15–30	3:1
Granodiorite gneiss	30–50	8–30	10–25	10–15	3:1

* Approximate.

GRANODIORITE GNEISS

Field relationships.—Outcrops of granodiorite gneiss occur along the Clearwater River from near the mouth of Jim Ford Creek to Ahsahka, along the North Fork from Ahsahka to near the mouth of Freeman Creek, and along Orofino Creek from Orofino to Whiskey Creek (Fig. 1). Quartz veins and pegmatites composed almost entirely of feldspar and quartz are numerous in the granodiorite gneiss, which grades into hornblende–quartz diorite gneiss and biotite–quartz diorite gneiss, with accompanying increases in mafic minerals and plagioclase and a decrease in potash feldspar. Most of the marble of the entire area is interbedded with granodiorite gneiss.

Megascopic description.—The color of the granodiorite gneiss is light to moderate gray, and the granularity is fine to coarse. The rock shows rather distinct light and dark bands, the difference being due to different proportions of feldspar and quartz to biotite and hornblende. Schistose and massive rocks of granodi-

orite composition occur locally in the gneiss but form only a minor portion of the rock as a whole and, consequently, are not shown separately on the map (Fig. 1). In the field the rock is almost indistinguishable from some phases of both the hornblende–quartz diorite gneiss and the biotite–quartz diorite gneiss, the difference being evident only upon microscopic examination.

Microscopic description.—The granularity of the rock is mainly fine to coarse. Some of the larger grains, which are porphyroblastic in origin, are 6–8 mm. in diameter. The actual fineness of grain in places appears to be somewhat accentuated by the manner in which porphyroblasts of plagioclase and microcline penetrate into and between fine-grained earlier minerals, giving the appearance of crushing. None of the major minerals displays euhedral crystal outlines; instead, they are embayed and irregularly penetrated by adjacent minerals.

The texture and the habit of the minerals of this rock are much like those of the hornblende–quartz diorite gneiss and the biotite–quartz diorite gneiss. The differences in the rocks lie mainly in the minerals present and in the proportions of minerals which are common to all three rocks. These differences are shown in Table 1.

The minerals which form the greater part of the granodiorite gneiss are quartz, biotite, hornblende, plagioclase, and microcline. Other mineral constituents which occur in variable and generally minor amounts are pyroxene, uralite, scapolite, allanite, epidote, zoisite, clinozoisite, chlorite, apatite, sphene, zircon, garnet, pyrite, magnetite, sericite, and carbonate.

Early quartz, early biotite, and early andesine in the granodiorite gneiss have properties, habits, and spatial relationships almost identical with those described for the same minerals in the hornblende–quartz diorite gneiss. They occur with the same suite of minerals as is present in the latter rock. Furthermore, they are associated with pyroxene, scapolite, and microcline, which additional associations will be discussed in the descriptions of these minerals.

Pyroxene is rare and occurs only locally. The pyroxene cleavage and the extinction angle of 42°, measured between the cleavage traces in longitudinal section and the *Z*-vibration direction, together with the pale-green color, the rather high relief, and the fairly strong birefringence, show it to be diopside. It forms somewhat blocky to irregular grains, from which prongs have grown between and around the early quartz and the early andesine, completely isolating some of the individual grains. A brown, very finely fibrous mineral, which was identified as uralite, grows along fractures and penetrates into the diopside perpendicular to the fracture surfaces. In some grains small, light-green, pleochroic, spearhead-shaped crystals seem to develop from the uralite and to form hornblende in the diopside. Other diopside grains are penetrated by well-developed prismatic crystals of hornblende, which do not appear to be associated with the uralite. These relationships suggest that the pyroxene is intermediate in age between early andesine and hornblende.

The hornblende is like that previously described in the other metasedimentary gneisses, except for its relationship to diopside and uralite.

Late andesine, although of about the same composition and physical and optical properties as the late andesine of the hornblende–quartz diorite gneiss, has

undergone, as has the early andesine, certain modifications which seem to be associated with the development of microcline. These changes will be presented with the description of the latter mineral.

The properties and spatial relationships of second-generation biotite are essentially the same as those previously described in the other metasedimentary gneisses. The proportion of biotite to hornblende is greater than that in the hornblende–quartz diorite gneiss but is about the same as that in the biotite–quartz diorite gneiss (see Table 1).

Scapolite occurs in some of the granodiorite gneiss which is adjacent to or near marble. The grains of this mineral are larger than the average grains of the rock. They show low relief, moderate birefringence, good cleavages in two directions at right angles to each other, and generally good uniaxial negative interference figures. Scapolite seems to embay diopside and late andesine crystals and, in turn, is penetrated by vermicular growths of clinozoisite. Age relationships with other minerals are not distinct, but the relationships just described indicate that scapolite is younger than late andesine and older than clinozoisite.

The relative age of the minerals of the epidote group to the other minerals and the properties and habits of all except zoisite are like those previously described in the hornblende–quartz diorite gneiss. Zoisite is especially abundant in rocks near or adjacent to marble layers. It has peculiar bluish-white to gray interference colors, and many grains do not become entirely extinct in any position. It forms irregular, coarse, lattice-like growths, with polysynthetically twinned feldspar. Small thin stringers of zoisite pierce diopside, late andesine, and scapolite.

Microcline forms 8–30 per cent of the rock. It occurs in anhedral grains of variable size, which show the gridiron structure and the undulatory extinction common to the mineral. Microcline penetrates irregularly into and between all the minerals described previously under this rock type and in places surrounds them, leaving isolated remnants of these minerals as islands in microcline. Where microcline replaces andesine, myrmekitic quartz develops in the andesine (Pl. II, figs. 1–3). Myrmekitic growths of quartz occur in all the rocks which contain microcline, and all but a few andesine grains which contain myrmekite are definitely in contact with microcline. Even those few grains are in contact with a mineral identified as feldspar and may be either microcline or orthoclase. Apparently, as the microcline replaced andesine, quartz, in an accompanying process, formed vermiform and rodlike bodies in the andesine but not in the microcline. In some corroded grains of andesine the myrmekitic quartz projects into the microcline (Pl. II, fig. 3). This condition is brought about by continued penetration and corrosion of the andesine by the microcline, which causes the myrmekitic quartz, which is replaced less readily than the andesine, if at all, to become partially or wholly enclosed in the microcline.

In addition to the early quartz and the myrmekitic quartz, there is a third generation of this mineral, which occurs as small, irregular veinlets and lobate grains and which corresponds to the late quartz in the previously described metasedimentary gneisses. This quartz displays the same optical properties and spatial relations as does the late quartz of those rocks (Pl. II, fig. 1).

Apatite, sphene, and zircon are more abundant in this rock than in the hornblende–quartz diorite gneiss. These minerals commonly occur as small grains

along cracks in the late-quartz veinlets, in addition to occurring in intergranular spaces and in cracks in other minerals. Apatite and zircon, especially, are commonly surrounded by fringelike borders of magnetite, which somewhat destroy the crystal outlines of the surrounded minerals (Pl. II, fig. 4).

Garnet and pyrite occur locally and sparingly, being commonly associated with late quartz. Some of the garnet is in large grains, which enclose ragged inclusions of biotite and hornblende. Small garnet and pyrite grains are subhedral.

The occurrence of magnetite is the same as it is in the other metasedimentary rocks (Pl. II, fig. 4), and it is considered to be the last mineral constituent formed, with the possible exception of sericite and carbonate.

MARBLE

Field relationships.—Layers of crystalline limestone, or marble, ranging in thickness from 6 to 600 feet, occur intercalated with the gneisses. Most of these layers are interbedded with the granodiorite gneiss in a limited area around Orofino, extending from near Jim Ford Creek, south of Orofino, to about $1\frac{1}{2}$ miles north of Ahsahka. These layers strike in a general northwest-southeast direction more or less parallel to the Clearwater River and dip mainly to the northeast. Outside the area of granodiorite gneiss, about 11 miles east of Orofino at Lime Mountain, a layer of marble about 600 feet thick is interbedded with biotite–quartz diorite gneiss. The marble at this location has undergone more intense silication than that at the other locations described.

Megascopic description.—The marble is white to bluish-gray, coarsely crystalline, and massive. Chemical analyses[7] show that some of the rock is almost entirely calcium carbonate. However, some specimens studied by the writer effervesce only weakly in lump form with cold dilute hydrochloric acid, indicating a fairly high percentage of magnesia. In some places the rock also contains pyroxene, amphibole, feldspar, sphene, magnetite, pyrite, and graphite, all discernible megascopically.

Microscopic description.—The purer rock is almost 100 per cent calcite and dolomite, but it also contains small amounts of talc, quartz, apatite, magnetite, graphite, and pyrite. The silicated rock contains, in addition, diopside, sphene, and zircon.

The carbonate of this rock shows curved twinning lines, which almost obliterate the cleavage traces. The only definite relation observed between cleavage and twinning was that of dolomite twinning parallel to both the long and the short diagonals of cleavage rhombs, seen in a thin section from a specimen, which in lump form did not effervesce freely in cold dilute hydrochloric acid. However, other specimens did react readily to acid, which fact, with the chemical analyses, shows that they contain a high percentage of calcite. In the silicated marble, isolated grains of carbonate, some of which may be secondary in origin, occur in diopside, tremolite, and plagioclase. Veinlets of secondary carbonate fill cracks in other minerals and in carbonate.

Locally, pyroxene may form as much as 80 per cent of the rock. It occurs in large subhedral crystals, which attain a length of 5 mm. The mineral shows good pyroxene cleavage, grayish-green color, rather high relief, fairly strong birefringence, and extinction angles measured to Z of about 42°, all of which are characteristic of diopside. This mineral grows into and around carbonate, leaving small grains scattered through the diop-

[7] A. L. Anderson, pp. 56 and 58 of ftn. 1.

side crystals. Isolated parts of grains of diopside in optical continuity occur in tremolite, plagioclase, and secondary carbonate.

Tremolite forms only a minor part of the rock and generally occurs in small prismatic crystals with diamond-shaped cross-sections, which penetrate diopside and carbonate and form acicular projections along cracks and cleavage directions in these minerals. These spatial relationships indicate that the tremolite replaces the diopside and some of the carbonate and is, therefore, a later mineral than they. Small, irregular, isolated grains occur in plagioclase and secondary carbonate.

Plagioclase is present in subhedral crystals, which vary in size from less than 0.25 to 5 mm. across. Most of the crystals show good polysynthetic twinning, and some of them are also twinned according to the Carlsbad law. Small grains of plagioclase are engulfed by larger crystals of plagioclase, suggesting two stages of development. The smaller, enclosed grains are calcic oligoclase, about $Ab_{72}An_{28}$, and the larger crystals are sodic andesine, about $Ab_{68}An_{32}$. Paragenetic relationships between oligoclase and other minerals are obscure, but tremolite seems to cut across and to grow around small oligoclase grains, whereas andesine embays and encloses carbonate, diopside, and tremolite and forms fine fingers along cracks and cleavage directions in the latter two minerals.

A peculiar mineral, which seems to form by the alteration of plagioclase, was identified as zoisite. This identification was based on the association with andesine, the lack of color, the weak birefringence with grayish-green interference colors, and the high relief relative to the adjacent andesine. Because of the habit of the mineral, it was not possible to determine its cleavage, extinction angle, optic character, or axial angle. It occurs in large, irregular, skeleton-like crystals lying in the plagioclase grains. This relationship indicates that zoisite is younger than andesine and has replaced that mineral.

Chlorite occurs sparingly, mainly as fibrolamellar grains in irregular to sheaf-like masses, which seem to grow into the sides of plagioclase crystals. No relationships between chlorite and tremolite or diopside were noted.

Talc occurs in small quantities. In the high-grade marble it grows along cracks in, and sends minute stringers into, the carbonate grains. Some talc is surrounded by vein quartz in the cracks. In the silicated marble, talc penetrates along cracks and cleavage directions in diopside and tremolite.

Quartz forms veinlets and lobate bodies in and between the minerals described above. These bodies penetrate and embay carbonate, diopside, tremolite, and plagioclase. Quartz is not abundant in any of the marble examined.

Apatite and magnetite occur in minor amounts in all the sections studied. They grow along intergranular cracks and penetrate into the bordering minerals in the manners described in the gneisses. Magnetite appears to be the last mineral formed.

In this rock type, sphene and zircon occur only in the silicated marble. Graphite and pyrite form small crystals in the high-grade marble. All four of these minerals are rare, and their paragenetic relationships are obscure.

Petrogenesis

Metasedimentary rocks

The bedding occurring in distinct, thick to thin layers, as in sedimentary rocks; the gradations into quartzites, on the one hand, and into igneous rocks, on

the other; and the mosaics of granular interlocking grains of early quartz indicate that the present gneisses were originally quartzites which have been modified by magmatic solutions. Because of the close proximity of the Idaho batholith, the gradational contacts of the gneisses with the batholith, and the mineralogic similarity between the batholithic rock and the gneisses, it is believed that these solutions came from the Idaho batholith, which is exposed east of the metamorphic zone and which probably lies at rather shallow depths below the metamorphic rocks as well. The sequence of mineral deposition indicates that the temperature at which mineral formation occurred increased as the invaded rocks were heated, reached a maximum, and then decreased normally. The action of these solutions on the siliceous sediments is thought to have produced the following changes:

1. Biotite was formed early in the metamorphic process while the rocks were under the stress from the intrusion of the batholith, which condition resulted in the alignment of the biotite flakes. The biotite is considered to have been added and not to have been formed from original constituents in the sediments. This premise is based on the quartzitic character of the quartz and the replacement nature of the biotite and on observations by A. L. Anderson[8] and A. L. Anderson and V. Hammerand,[9] who found in comparable areas of these rocks that biotite occurs in the contact zones but not in the unaltered quartzites.

2. As the temperature of the rocks was raised and the stress relieved, sodic andesine was deposited, partially replacing early quartz and early biotite.

3. When the temperature reached its maximum, diopside was formed locally. Apparently, this condition did not prevail long, because very little diopside was formed except in some of the marble layers.

4. After this high temperature was reached, the solutions became cooler, and some of the diopside was partially altered to uralite.

5. Hornblende formed at the expense of all the minerals which preceded it. The amount of hornblende deposited varied with the access, the amount, and the composition of the feeding solutions. Locally, copious solutions rich in basic materials invaded the rocks and formed coarse-grained hornblendite. According to A. L. Anderson,[10] these solutions were "rendered highly fluid by dissolved mineralizers."

6. Late porphyroblastic andesine penetrated, molded itself around, and isolated previously formed minerals.

7. A second generation of biotite was formed. The relationship of this biotite to hornblende and late andesine proves that it is different from the early biotite. There is some indication that late biotite forms the fine, wisplike projections on the early biotite flakes. J. L. Gillson,[11] in describing a quartz monzonite in the Pioche District, Nevada, mentions such a late growth on biotite. A. L. Anderson and V. Hammerand[12] and A. L. Anderson[13] describe similar growths which penetrate andesine crystals. Both early and late biotite vary in amount from place to

[8] "Contact Phenomena Associated with the Cassia Batholith, Idaho," *Jour. Geol.*, Vol. XLII (1934), p. 386.

[9] "Contact and Endomorphic Phenomena Associated with a Part of the Idaho Batholith," *Jour. Geol.*, Vol. XLVIII (1940), pp. 566–67.

[10] P. 96 of ftn. 6.

[11] "Petrography of the Pioche District, Lincoln County, Nevada," *U.S. Geol. Surv. Prof. Paper 158-D* (1929), p. 88.

[12] Pp. 571 and 574 of ftn. 9.

[13] P. 1105 of ftn. 5.

place, with variations in the quantity and character of the emanations.

8. Scapolite formed locally, replacing diopside and late-andesine crystals.

9. As the solutions penetrated between andesine crystals and along cracks in the andesine, they altered this mineral to a more sodic plagioclase and minerals of the epidote group. As further alteration occurred, some of these products probably were partially dissolved, transported in solution, and redeposited as minerals of the epidote group not associated with andesine but as replacements in hornblende and biotite. It is also possible that these minerals may in part have been formed as alteration products in hornblende.

10. Microcline was deposited in certain restricted areas, indicating further variations in the composition of the emanations given off by the batholith. This mineral penetrates, lies between, and engulfs those minerals which precede it. As andesine was replaced by the microcline, myrmekitic quartz was developed as an accompanying process in the andesine, but in no other mineral. On continued replacement of the andesine, myrmekitic quartz, which is apparently less or not at all susceptible to replacement by the microcline, became wholly or partially surrounded by the latter mineral.

11. Late quartz formed veinlets and lobate bodies, which filled interstitial spaces and replaced adjacent minerals.

12. Apatite, sphene, and zircon apparently followed late quartz, since they occur in cracks in this mineral. Apatite and sphene, and possibly zircon, also replaced andesine, hornblende, and biotite.

13. Magnetite developed along cracks between and penetrated the previously formed minerals.

14. Sericite was formed in the feldspar and carbonate in many of the minerals.

The habits and relative scarcity of garnet, pyrite, and graphite do not permit accurate determination of their paragenetic positions.

Since all the minerals in the gneisses except original quartz were introduced, the solutions which introduced them were of necessity complex. Furthermore, the solutions probably were not always at the same temperature or of the same composition from place to place or from time to time during the period of metamorphism. Stress conditions disappeared or became minor after the biotite stage.

IGNEOUS ROCKS

ULTRA-BASIC ROCK—DUNITE

The minerals of this rock were formed by deposition from, and subsequent reaction by, a cooling magma and its separating solutions. The steps in this process which produced the rock in its present state were as follows:

1. Olivine was deposited from the cooling magma.

2. As the solutions cooled, the amphiboles replaced the olivine. Tremolite seems to have formed earlier than anthophyllite, but their relationships are not distinct, and there may be an overlapping in time of formation of the two minerals.

3. Chlorite replaced amphibole and formed small, irregular, or sheaflike masses in the amphiboles.

4. Talc formed by replacing olivine and amphibole. It may be later than chlorite, or the two minerals may be partially or wholly contemporaneous; but the habits of the two minerals are so similar and their mutual boundaries have been so destroyed by the growth of serpentine that accurate relationships between them cannot be determined.

5. Iddingsite formed around the periph-

eries and in cracks in the olivine, penetrating from the cracks into the grains. The exact paragenetic position of the iddingsite is uncertain, but its high water content indicates that it belongs close to this position in this series.

6. Rather abundant serpentine formed in most of the rocks, replacing olivine, amphibole, chlorite, talc, and iddingsite.

7. Quartz, apatite, magnetite, and carbonate were deposited late, the quartz forming veinlets between most of the other minerals.

The solutions which caused the changes in the dunite are believed to have been simple, cooling solutions and not the complex ones which were required to form the minerals in the gneisses. This is indicated by the normal sequence of mineral formation from olivine to serpentine, the similarity in chemical composition of the minerals, and the lack of parallelism of these minerals to those in the surrounding gneisses either in variety or in temperature of formation. It is believed, therefore, that the rock was altered by autometamorphic, or dueteric, processes in the sense defined by J. J. Sederholm.[14]

Conclusions

1. These complex, hornblende-rich rocks are substantially the same in character and origin, except for the high hornblende content, as the other gneisses of this general region which were formed by granitization.

2. The evidence that the metamorphism described in this paper was caused by hydrothermal solutions is as follows:

a) The gradations of the gneisses into unquestionable quartzites, on the one hand, and into igneous rocks, on the other

b) The fact that thin-section study shows interlocking grains of quartz, which are evidently remnants of quartzite, replaced by a complex suite of minerals common to igneous rocks

c) The sequence, the over-all chemical complexity, and the replacement relations of the constituent minerals of these rocks

d) The mineralogic similarity between the gneisses and the rocks of the Idaho batholith

e) The exceptionally large size of the hornblende crystals which occur in the hornblendite

3. The variations in composition of all the rocks of this area except dunite—as regards, among other minerals, hornblende, biotite, and microcline—are the result of diversity in the structure of the intruded rocks, of variations in the depth of the batholith below these rocks, and of variations in the composition and the quantity of the solutions given off by the batholith from place to place and from time to time.

4. The myrmekite, in these rocks, is the result of replacement of andesine by microcline, with the accompanying development of vermicular quartz in the andesine.

5. The solutions which caused the granitization were complex, hydrothermal solutions. The temperatures of mineral deposition rose as the intruded rocks became heated and then gradually fell to normal levels. On the other hand, the alteration of the ultra-basic intrusives was caused by simple, cooling solutions by

[14] "On Synantetic Minerals and Related Phenomena," *Comm. Geol. Finlande Bull. 48* (1916), pp. 4 and 142.

autometamorphic, or deuteric,[15] processes.

6. The granitized rocks were under stress during the early stages of granitization, as indicated by the alignment of the biotite.

7. Because of the differences in the character of the solutions which formed the gneisses and the dunite, the latter rock is thought to be different in age from the gneisses; and, because the dunite was not affected by the granitizing solutions, it may well be younger than the gneisses.

8. The hornblendite is thought to represent a special phase of the granitization process because of the similar mineralogy of the hornblendite to that of the gneisses, differing in proportion rather than in character of minerals, and because of its gradational borders into the gneisses.

9. The extremely coarse texture and the high concentration of basic minerals in the hornblendite lead to the conclusion that it was formed by profuse aqueous solutions.

[15] These two terms are here used in the sense defined by Sederholm (*ibid.*).

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THE COMPOSITION OF METEORITIC MATTER

II. THE COMPOSITION OF IRON METEORITES AND OF THE METAL PHASE OF STONY METEORITES

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ABSTRACT

The average composition, with respect to iron, nickel, and cobalt, of iron meteorites and the metal of stony meteorites has been determined by statistical study of the analyses of 320 meteorites. The average composition of iron meteorites is: Fe, 90.78 ± 0.26 per cent; Ni, 8.59 ± 0.24 per cent; Co, 0.631 ± 0.019 per cent. The average composition of the metal phase of stony meteorites is: Fe, 88.58 ± 0.55 per cent; Ni, 10.69 ± 0.51 per cent; Co, 0.705 ± 0.056 per cent. It is shown that the higher values of nickel and cobalt in the metal phase of stony meteorites result from a relationship whereby the nickel and cobalt concentrations in the metal phase increase as the metal-phase content of the meteorite decreases.

I. INTRODUCTION

In a recent paper,[1] the authors presented the results of a statistical study of a number of analyses of the silicate phase of stony meteorites. The study has been extended to iron meteorites and to the metal phase of stony meteorites. Averages for the composition of iron meteorites have previously been given by O. C. Farrington[2] and by I. and W. Noddack.[3]

The present study was undertaken for two purposes: first, to determine the accuracy with which one might express the average chemical composition of both iron meteorites and the metal phase of stony meteorites and, second, to determine whether or not significant differences in average chemical composition exist between these substances.

II. METHOD AND RESULTS

From the data available on iron meteorites, 220 analyses were selected in which the analyses for iron, nickel, and cobalt appeared to be reasonably satisfactory. Eleven falls and 209 finds were used, of which 151 have been listed by O. C. Farrington,[4] 33 by C. A. Doelter,[5] and 36 have come from more recent sources. In view of the general inadequacy of the analyses for the minor constituents and also in view of the fact that the minor constituents comprise but a small fraction of the whole, the study embraced only the major constituents—iron, nickel, and cobalt. In all cases the sum of these three elements was adjusted to 100 per cent, and the statistical analysis was then made on that basis.

Frequency curves showing the occurrence of iron, nickel, and cobalt in iron meteorites are shown in the upper half of Figure 1. Abundance averages for these elements, together with standard deviations and precisions,[6] are given in Table 1.

One hundred analyses of stony meteorites were accumulated, over half of them falls, in which the metal phase had

[1] "The Composition of Meteoritic Matter. I. The Composition of the Silicate Phase of Stony Meteorites," *Jour. Geol.*, Vol. LV (1947), pp. 405–11.

[2] "Analyses of Stone Meteorites," *Field Mus. Nat. Hist. Geol.*, Vol. III (1911), p. 212.

[3] Die Häufigkeit der chemischen Elemente," *Naturwiss.*, Vol. XXXV (1930), p. 757.

[4] "Analyses of Iron Meteorites Compiled and Classified," *Field Mus. Nat. Hist. Geol.*, Vol. III (1907), p. 59.

[5] *Handbuch der Mineralchemie*, Vol. III, Part II (1926), pp. 574–607.

[6] See ftn. 1.

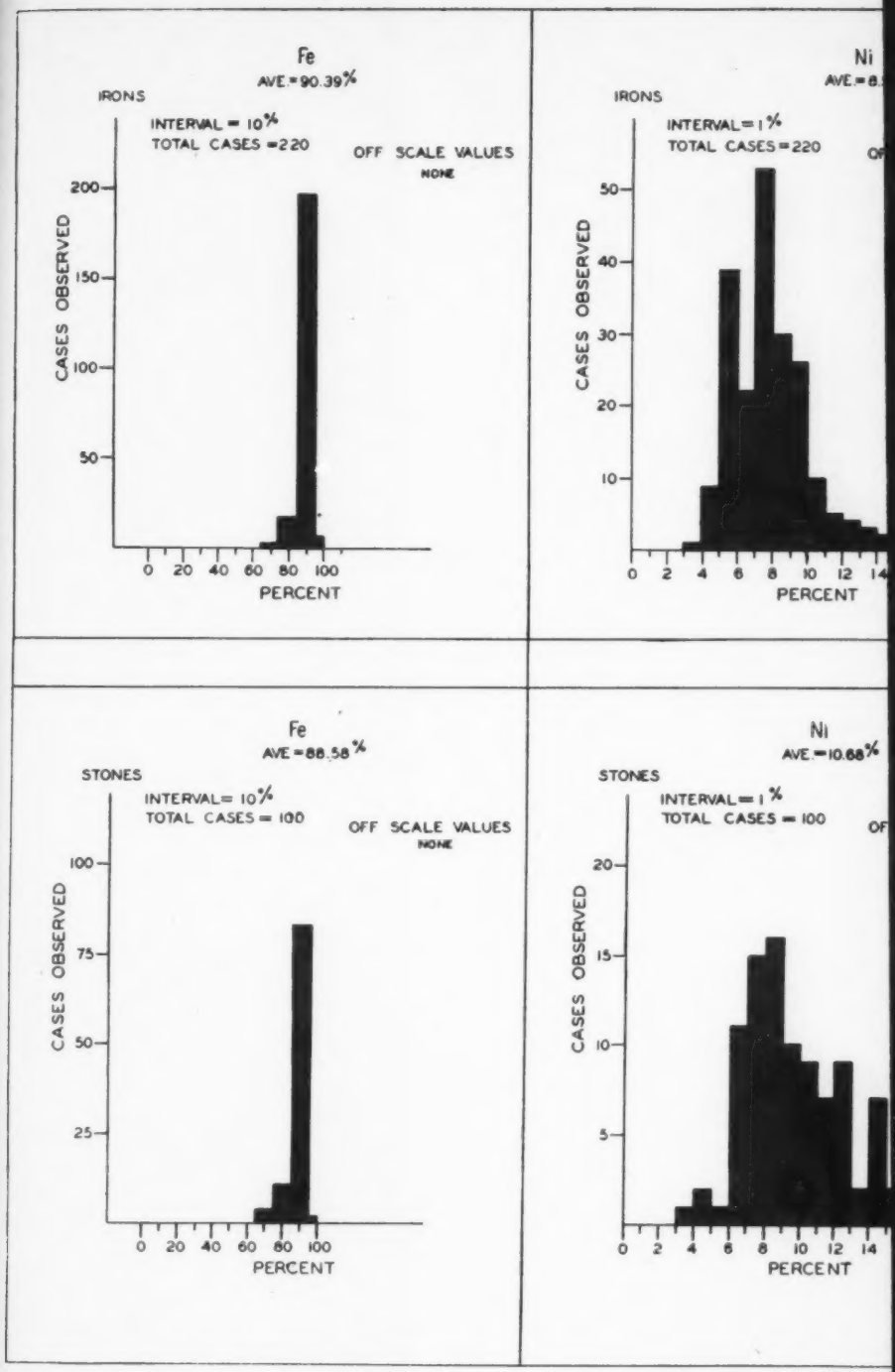
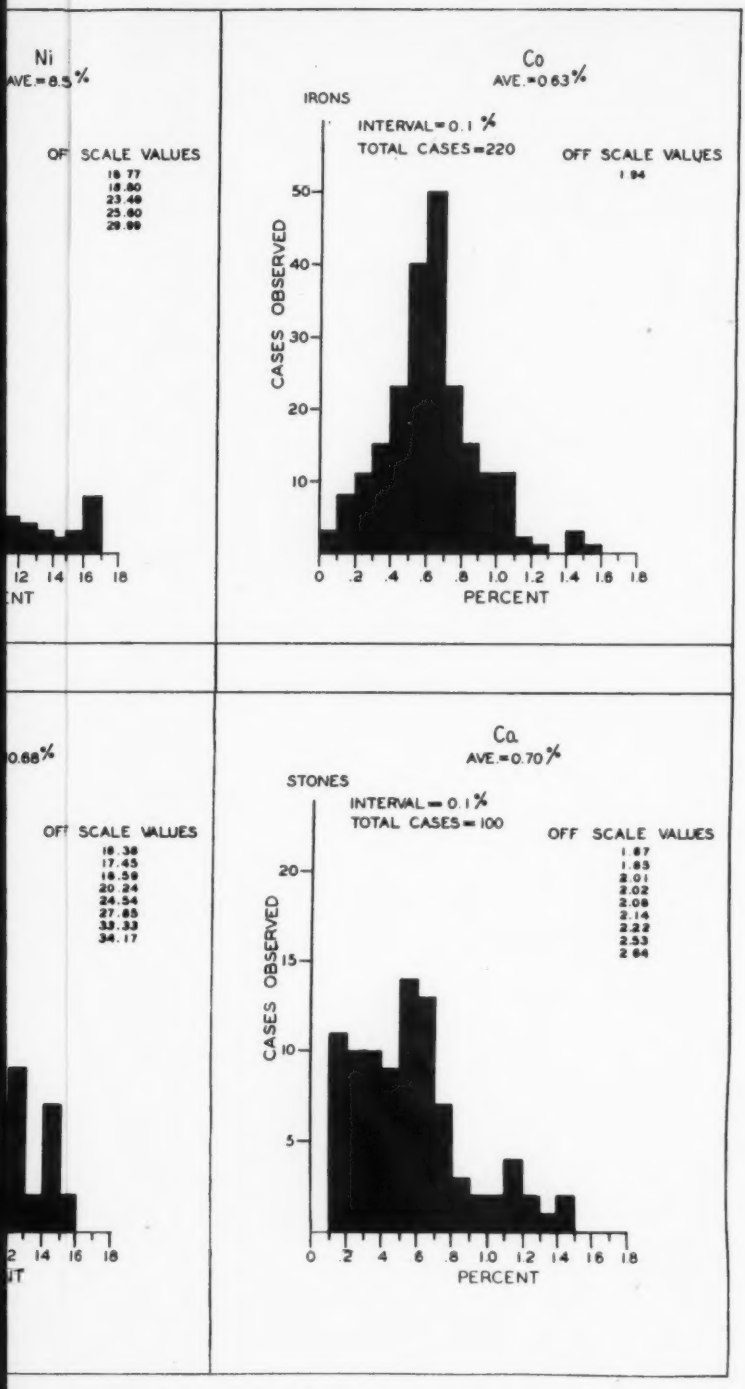


FIG. 1.—The frequency of occurrence of iron, nickel, and cobalt in iron meteorite



eteorites, compared with the metal phase of stony meteorites

been analyzed in an adequate manner. Of the stony meteorites used, 71 are listed in the authors' previous paper,[6] and 29 are from other sources. As with the iron meteorites, only data for iron, nickel, and cobalt were used, the sum of these three constituents being adjusted to 100 per cent.

Frequency curves showing the occurrence of iron, nickel, and cobalt in the metal phase are shown in the lower half of Figure 1. Abundance averages for

TABLE 1

THE RELATIVE ABUNDANCE OF IRON, NICKEL, AND COBALT IN IRON METEORITES AND IN THE METAL PHASE OF STONY METEORITES

	Per Cent Fe	Standard Deviation	Per Cent Ni	Standard Deviation	Per Cent Co	Standard Deviation	Total Cases
Metal phase of stony meteorites	88.58±0.55	5.46	10.69±0.51	5.11	0.705±0.056	0.555	100
Iron meteorites	90.78±0.26	3.84	8.59±0.24	3.59	0.631±0.019	0.277	220
Average of Farrington* for iron meteorites	90.85	†	8.52	†	0.59	†	318
Average of Noddacks‡ for iron meteorites	90.2	†	8.46	†	0.547	†	16

* See ftn. 4. † Not determined. ‡ See ftn. 3.

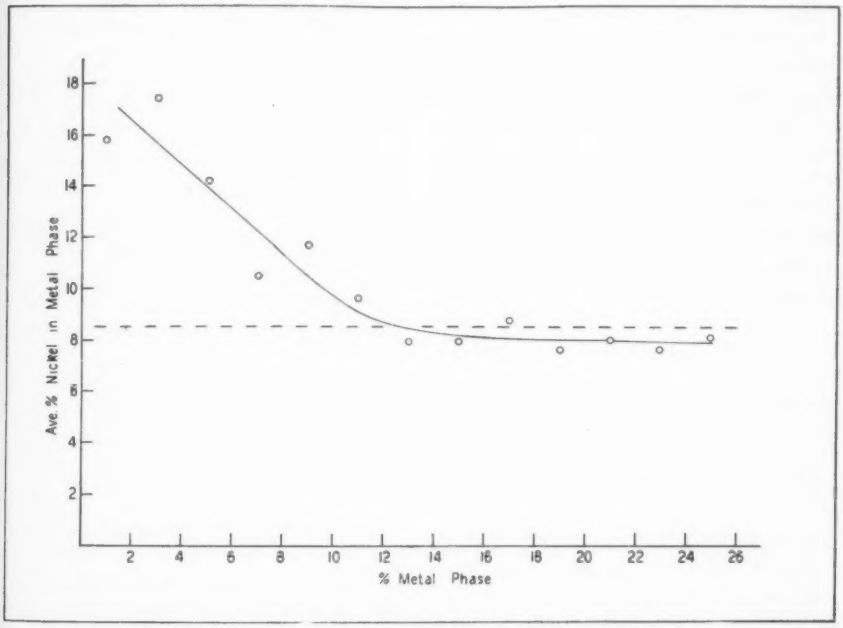


FIG. 2.—The variation of the nickel content in the metal phase of stony meteorites with the metal-phase content in the meteorite. The dotted line represents the average nickel content of iron meteorites. Each point represents the average of between 2 and 19 stony meteorites.

these elements, together with standard deviations and precisions, are given in Table 1.

III. COMPARISON OF RESULTS

A comparison of the figures shown in Table 1 indicates the following facts:

a) The average nickel content of the metal phase of stony meteorites is greater than the average nickel content of iron meteorites.

b) Although the effect is less pronounced, the average cobalt content of the metal phase of stony meteorites is apparently greater than the cobalt content of iron meteorites.

c) As is shown by the standard deviations and by the frequency curves, the spreads in concentration of iron, nickel, and cobalt are greater in the metal phase of stony meteorites than in iron meteorites.

A more detailed study of the data reveals that the higher average nickel and cobalt concentrations associated with the metal phase of stony meteorites results from a pronounced relationship between the nickel concentration in the metal phase and the total metal-phase concentration in the meteorite. If one arranges a statistical sampling of stony meteorites in the order of increasing metal-phase concentration and averages the nickel concentrations over specified intervals of metal-phase concentrations, the curve shown in Figure 2 is obtained. It can be seen from Figure 2 that, as the concentration of metal phase associated with meteorites decreases, the nickel concentration in the phase increases to a value that is approximately double the average nickel concentration in iron meteorites, represented by the broken line. The thirteen points defining the curve represent averages calculated from 96 stony meteorites, of which 60 had a metal-phase content of less than 10 per cent.

The rapid rise in nickel concentration with decreasing metal-phase concentration begins at a metal-phase concentration of approximately 13 per cent. At metal-phase concentrations greater than 13 per cent, the nickel concentration levels off to a value slightly lower than, but nevertheless very close to, the average nickel concentration in iron meteorites.

These latter observations substantiate conclusions to the same effect drawn previously by G. T. Prior.[7]

[7] "On the Genetic Relationships and Classification of Meteorites," *Min. Mag.*, Vol. XVIII (1916), p. 26; and "The Classification of Meteorites," *ibid.*, Vol. XIX (1920), p. 54.

UPRIGHT TRUNKS OF *NEOCALAMITES* FROM THE UPPER TRIASSIC OF WESTERN COLORADO

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ABSTRACT

This article describes a hitherto unreported find of *Neocalamites* from the Upper Triassic of western Colorado. The find is of unusual interest because of the large size of the trees and their geologic mode of occurrence. Moreover, the large number of plants exposed indicates the presence of an Upper Triassic forest of considerable size.

INTRODUCTION

Many fossils and something of the woody structure and life-history of the Equisetales are fairly well known from the Devonian to the end of the Paleozoic, when the well-known genus *Calamites* became extinct. But our knowledge of the history of this great group of plants after the end of the Paleozoic is rather meager. That the group survived the Paleozoic is known from scattered finds reported from time to time from the Mesozoic and from the single present genus *Equisetum*. These modern horsetails are the only living representatives of a much greater group of plants that dominated the Carboniferous landscape.

The *Calamites* of the Carboniferous grew to great size. Trunks 30 feet long and 1 foot in diameter have been reported. The number of individuals must have been very great, for their casts and impressions are found in large numbers in Upper Paleozoic rocks.

When we come to the Mesozoic, however, the Equisetales, though still found, are rare, and those that have been reported are usually of small size compared with their Paleozoic predecessors.

The find described here from the Upper Triassic of western Colorado is therefore of more than passing interest, not only because of its rarity but also because of the large size of the fossil trunks. Furthermore, the plants are all still in the vertical position of growth, and consequently have not been compressed or distorted.

These fossils were first discovered by the writer in the spring of 1942, but further study was interrupted by army service until 1946, when another trip was made into the area. The fossil locality is in west-central Colorado, on the south side of the Dolores River and about a mile from the town of Bedrock, Colorado.

Preliminary examination indicates that the forest here discovered is comparable in many respects to that of the celebrated Joggins, Nova Scotia, section of Carboniferous age, where upright trunks of *Calamites* and *Sigillaria* have excited the admiration of geologists since the days of Lyell, Logan, and Dawson. According to Charles Schuchert and C. O. Dunbar,[1] some of the trees at Joggins have a length of 20 feet, and a few are 4 feet in diameter. While these trees from Joggins are much thicker than those described here, it is believed that these Triassic plants were fully as long. In both localities, however, the tops have been broken off, thus making it impossible to

[1] *A Textbook of Geology*, Part II, "Historical Geology" (New York: John Wiley & Sons, Inc., 1933), pp. 249–50.

get a true picture of the actual height. A comparison of the photographs of the Colorado plants with those shown by Schuchert and Dunbar[2] of the Joggins fossils shows a striking similarity.

STRATIGRAPHY AND MODE OF OCCURRENCE OF THE PLANTS

The Triassic formations of the Colorado Plateau province consist of a great series of red sandstones, shales, and conglomerates, mostly nonfossiliferous and probably in large part of continental origin. In Arizona and Utah these "Red Beds" have a threefold division, the Moenkopi, Shinarump, and Chinle. According to J. Gilluly and J. B. Reeside,[3] these strata are 2,232 feet thick in eastern Utah. However, the Triassic rapidly thins eastward into western Colorado; and, in view of this thinning to the east, it is doubtful whether either the Moenkopi or the Shinarump is represented in the fossil locality under discussion.

The Chinle formation of western Colorado, in which the fossils occur, is considered Upper Triassic in age. In this area it has a total thickness of about 300 feet and typically outcrops in fairly steep slopes at the base of the Wingate sandstone. Opinion as to the age of the Wingate sandstone has varied with different writers. Some have assigned it to the Triassic and others to the Jurassic. It is a great, cliff-making sandstone in the plateau country and stands out as a sheer cliff several hundred feet high in the canyons, while the Chinle forms steep slopes below it.

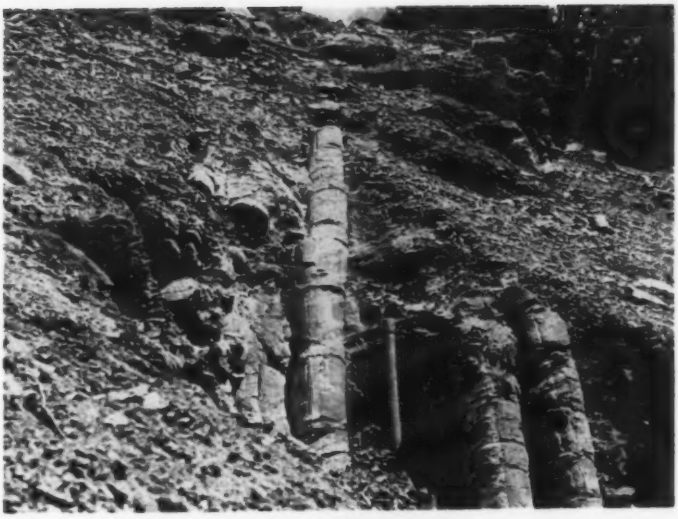
In this area the Chinle formation consists of micaceous red-brown sandstones, maroon sandy shales, and some thin limestone conglomerates. The sandstones are uniformly fine grained, few grains being large enough to be easily recognized. The red sands in which the fossils occur are also of very fine grain, and there is an absence of well-defined bedding planes. Two fossiliferous conglomerate layers, from 12 to 18 inches thick, containing *Unio* and bone fragments were found at the fossil-plant locality, one above the tops of the trees and the other just below the roots. These conglomerates carry limestone pebbles as much as $\frac{1}{2}$ inch in diameter, together with well-rounded quartz and feldspar fragments.

The fossil plants are embedded in this fine-grained sandstone between the two conglomerate layers and about 150 feet below the Wingate sandstone cliff. The trunks of the trees occur as casts, and some measure 12 inches in diameter. All trunks seen were *in situ* and in the vertical position of growth (Pl. I, *A*). They are quite close together. The trunks of five trees were fairly well exposed, and further search revealed the tops of many others projecting above the surface. A 15-foot section of one tree was removed, and some preliminary excavating was done in the hope of finding the roots, but this attempt was abandoned because of lack of time. Considering the 15-foot length of the specimen collected and the fact that the root was not reached and that the top was broken off, it is estimated that these giant Mesozoic Equisetales may have reached a height of well over 50 feet. The casts themselves are fairly well preserved and solid and show quite clearly the parallel grooves formed by the vascular bundles (Pl. II, *A*, *B*). The nodes and leaf scars are also well preserved, but no woody structure of the actual plant has yet been found. Although no leaf impressions were found, enough of the structure of the nodes is

[2] *Ibid.*

[3] "Sedimentary Rocks of the San Rafael Swell and Some Adjacent Areas in Eastern Utah," *U.S. Geo. Surv. Prof. Paper 150-D* (1928), p. 65.

PLATE I



A

B

A. View showing Neocalamites erect in the position of growth

B. Neocalamites *in situ*, western Colorado

PLATE II

Close-up of one trunk showing nodes and grooves made by the vascular bundles

preserved to identify the plants as *Neocalamites*.

L. H. Daugherty[4] has reported the presence of *Neocalamites* in the Upper Triassic of Arizona, but the plants that he describes are considerably smaller than these in Colorado. It is believed that the *Neocalamites* here described are the largest yet reported from the North American Triassic. Possibly these trees had reached their maximum size at that time, judging from the fact that those reported from higher horizons in the Mesozoic are, on the whole, noticeably smaller.

CONDITIONS OF SEDIMENTATION

Perhaps the most striking feature of the fossils is the evidence which they present of very rapid sedimentation. Although these hollow-stemmed plants, as now preserved, are about 20 feet high, yet they were rapidly and effectively buried before they could fall. The absence of coarse sediment and well-defined bedding planes in the strata apparently would indicate deposition under uniform conditions and probably by slow-flowing currents. Yet, if this is true, it is hard to see how slow-flowing currents could bury the plants before they fell. Possibly, sluggish, meandering streams, heavily laden with sediment, broke through natural levees of their channels onto the timbered low lands during the Chinle epoch of sedimentation and thus rapidly buried the trees.

Another problem is how the sands got into the hollow stems to form the casts. Two possibilities are suggested. First, the tops of the trees may have been broken off after rapid sedimentation had buried some 20 feet of their trunks, and sand then had access to the hollow interiors. A second possibility is that they were completely buried before the woody structure of the outside decomposed, thus allowing for the formation of molds, which were subsequently filled with sands. A careful search has revealed no trace of any woody structure; consequently, the writer is inclined to believe that the tops of the plants were broken off and the hollow interior subsequently filled with sand.

ACKNOWLEDGMENTS.—The writer wishes to thank Dr. John B. Reeside, Jr., and Dr. Ralph Brown, both of the United States Geological Survey, for reading the manuscript and making some helpful suggestions.

[4] "The Upper Triassic Flora of Arizona," *Carnegie Inst. Wash. Pub. No. 526* (Washington, D.C., 1941), p. 58.

THE RATE OF ROUNDING OF BEACH BOULDERS

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ABSTRACT

A demonstration is given of the ability of waves of moderate strength to round blocks of hard, fine-grained basalt as much as 2 feet in diameter in less than ten years. Mention is also made of an example of the short distance of travel that may be involved in the rounding of stream pebbles.

A little north of Takapuna Beach, which lies in Hauraki Gulf about 5 miles north of Auckland City, New Zealand, there has been demonstrated within recent years the rapidity with which waves of moderate strength are able to round blocks of hard rock. As quantitative data upon phenomena of this kind are relatively scarce, the writer has ventured to give the following account of the facts.

About twenty years ago the local borough council unwisely allowed the removal of a natural barrier of blocks of basalt which had long protected the shore from the sea waves at a locality indicated by the tip of the arrow on the map (Pl. I*A*). Here a subhorizontal sheet of basaltic lava, from 2 to 3 feet in depth, rests near high-water mark upon an unknown thickness of soft white "pipe clay" (Pl. I*B*), in some places with the intervention of a few inches of volcanic cinders. Where not highly vesicular, as it may be locally, the basalt, which is a hard, fine-grained, olivine-rich variety, is dissected by vertical joints into irregular polygonal columns usually from 18 inches to 2 feet across.

Upon removal of the blocks of basalt that had earlier formed the shore, the waves soon excavated the soft pipe clay from beneath the basalt, thereby undermining the latter and causing its collapse as a relatively unbroken sheet along a strip extending back from the earlier shoreline for about 6–8 yards. It was considerable time before the waves disintegrated this sheet into blocks sufficiently isolated to be subject to rolling one on another, and it is only within the last ten years that such rolling has been an effective process. During that time there has been continued, but slow, collapse of the edge of the sheet of basalt until its maximum recession is now probably about 15 yards, although the writer has no exact data upon this point. Plate II is a typical view (at low tide) of the boulders that strew the shore; many of the larger are still angular, but others of from 1 foot to over 18 inches in diameter are well rounded.

In comparison with exposed shorelines, the locality at which these boulders occur is sheltered, for it is protected from major swell by outlying islands, as is evident from the map (Pl. I*A*). The waters offshore are shallow, for at ½ mile from the shoreline the depth at mean low tide is a little over 3 fathoms; it increases to 5 fathoms at 1 mile and reaches its maximum of from 6 to 7 fathoms at 1½ miles. Although there is open water subject to frequent northeast gales for nearly 50 miles to the northeast, the boulder beach itself is not exposed directly to the waves generated by these gales, being protected by a short, low tongue of unbroken basaltic lava, which is shown in the background of Plate II.

Although waves have been stirring the boulders of this shore more or less effec-

PLATE I

A

B

A.—Map showing Hauraki Gulf near Auckland, New Zealand. The area described is at the tip of the arrow.

B.—Beach boulders and rubble lying on soft white "pipe clay" (foreground) at the shore near Takapuna, Auckland, New Zealand.

PLATE II

Boulders at the locality of Fig. 2. The largest spherical boulder measured is 1 ft. 10 in. in diameter.

tively for approximately ten years, it is only within the last four or five years that the result of their work in rolling, and thereby rounding, the large blocks of basalt has become obvious.

An observation pertaining to a related phenomenon was made on the rounding of stream pebbles in the vicinity of Whangarei, North Auckland. At McLeod's Bay, Whangarei Heads, a small stream rises in a sharp-crested range made of coarse-textured andesitic breccia and plunges down the precipitous face of the range for about $\frac{1}{4}$ mile before flowing on a moderate gradient to the shore at a little over $\frac{3}{4}$ mile from the divide. Where exposed by road cuts, its earlier fan-gravels include pebbles up to 2 inches in diameter, of which about 60 per cent show a surprisingly high degree of rounding. The hard, fine-grained andesite of which they are composed accepts a very smooth surface upon attrition, and a very large proportion of the rounded pebbles show perfection both of rounding and of smoothing. Prior to seeing this demonstration of the work of streams, the writer had imagined that far greater length of travel would be required before perfection of rounding could be attained; this misconception may be fairly general.

THE ORIGIN AND IMPORTANCE OF THE RALEIGH GRAPHITE

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ABSTRACT

There are several parallel tilted beds of earthy schistose graphite enclosed in the pre-Cambrian schists and gneisses of Wake County, North Carolina. Two or these beds are large enough to be of commercial interest and therefore were mapped along a 10-mile strike. From their structure and mineraliaztion it is evident that the graphite is part of a series of metamorphosed carbonaceous sediments. These beds will prove to be excellent key horizons for later work in the area.

INTRODUCTION

Schistose earthy graphite occurs in many beds in the pre-Cambrian schists and gneisses of Wake County, North Carolina. These graphite beds are grouped together and discussed in this paper as the "Raleigh graphite." All the exposures of the Raleigh graphite which were mapped lie in a belt 10 miles long and 2 miles wide. The northern end of the belt is at the town of Six Forks, nearly 10 miles north of Raleigh. How far south it extends has never been satisfactorily determined. Its southern end may be more than 18 miles southwest of Raleigh.[1] This area lies entirely in the Raleigh quadrangle of the United States Geologic Survey (Fig. 1).

DESCRIPTION OF THE GRAPHITE

The graphite is schistose with well-developed foliation. There is a variation in the appearance of this schistosity which is a direct reflection of the percentage of graphite in the rock. At most localities the schist averages about 4 per cent graphite. In this graphite-lean rock the foliation is extremely well developed; the schist is quite micaceous, light- to dark-gray, and quite hard. It will scratch a piece of paper rather than mark it, although black graphite may be rubbed from a hand specimen with the fingers. Locally, the graphite content of the schist may reach 35 per cent. If it is above 12 per cent, the schist is more massive, almost slaty, black, and very soft. Graphite of this type will mark a piece of paper easily. There is still a visible quantity of mica in the schist, although it may not show on the outside of a weathered lump. Specimens taken from old mine dumps appear to be pure graphite from the outside, as the mica has been washed off. A much better picture is obtained from the inside of these lumps, where the mica shows as minute, sparkling, contrasting flakes.

Quartz is a major constituent of both types of graphite schist and composes from 40 to 85 per cent of the rock's weight. Some of this quartz is plainly seen in veins and stringers. Most of it, however, is in a finely divided state and consequently difficult to see. It is also partly hidden by graphite, which has stained the minute cracks in its surface. Both types of schist, when ground to −200 mesh, almost invariably show quartz and graphite in every fragment. Garnet and kyanite occur in the graphite in a few places but are not common. These two minerals are localized in zones and are later than either the graphite or the schistosity.

[1] Ebenezer Emmons, *Geological Report of the Midland Counties of North Carolina* (New York and Raleigh, 1856), p. 221.

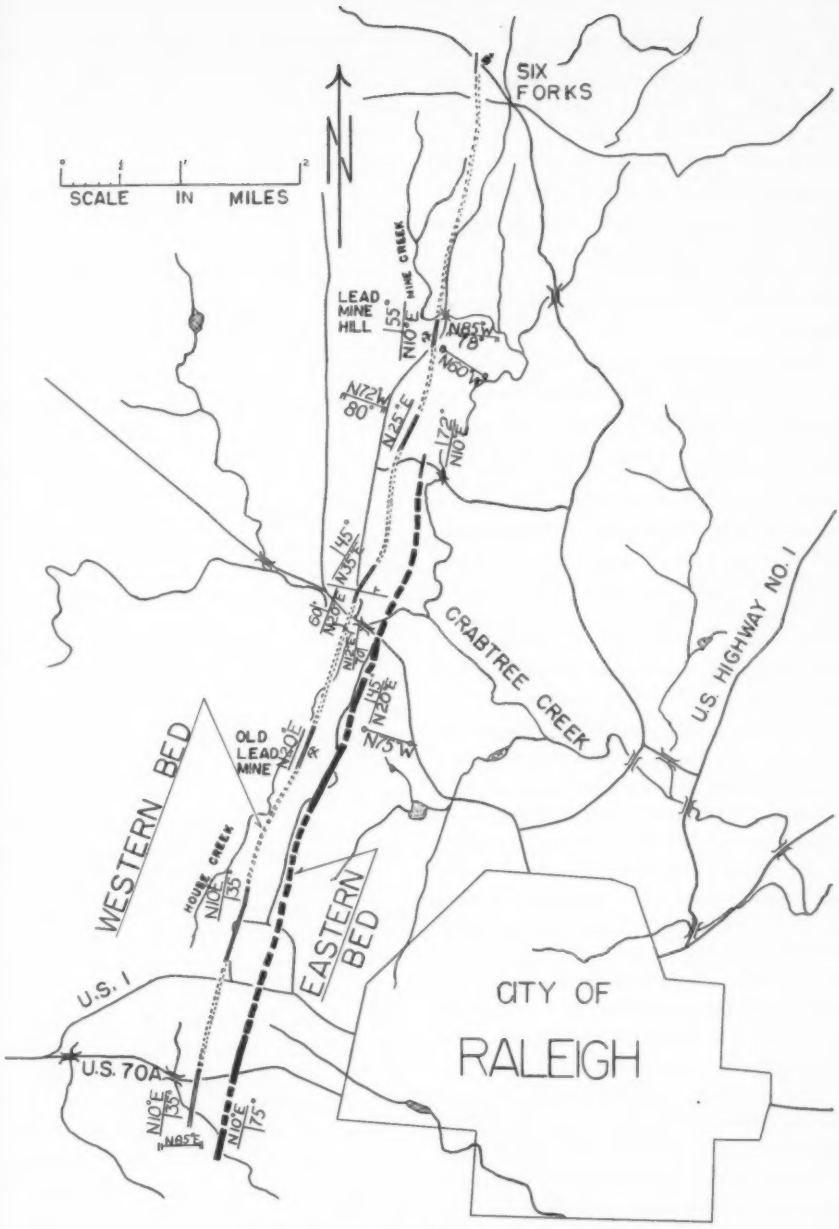

FIG. 1.—Geologic map of the Raleigh graphite

OCCURRENCE IN BEDS

It is difficult to select a term to describe the occurrence of the graphite which does not also predicate an origin. The term "bed" was selected, as it gives a clear picture of the interbedded appearance of the graphite layers and the schistosity of the enclosing gneisses and schists. Justification for this special usage is given in the section on the origin of the graphite.

Only two of the beds of graphite schist are wide enough to be considered economically important. These contain all the old mines and prospect holes, and they are the only beds mapped throughout the 10 miles. Both beds strike about N. 12° E. and are parallel to the schistosity of the country rock. A few miles north of Raleigh they curve slighly to the northeast. This change of strike direction is attributed to rotational faulting rather than to a simple bending of the regional schistosity.

There is but one fault plane exposed; however, other strike changes and outcrop locations indicate that there may be at least three more. This fault plane is clearly exposed in a road cut, striking N. 80° W. It is practically vertical. On the south side the graphite strikes N. 20° E. and dips 60° W. On the north side the same band strikes N. 35° E. and dips 45° W. There has been a torsional rotation of the graphite bed through an arc of 15°. The strike direction also changed 15°.

The geometry of such a fault is complicated beyond simple analysis by the degree of flexure, which may have been achieved before rupture took place and the fault plane itself developed. If a piece of paper is held so that it dips 60°, and then one end is twisted to give a change of dip to 45°, an unruptured flexure is formed. The strike changes along a horizontal section from the original strike at one end to a new strike at the other. Experiment with rigid and plastic models indicates that rupture could occur after torsion had produced the strike change. The angle of dip of the fault plane seems to be a function of brittleness. This type of faulting is unusual, and the author is not prepared to give a quantitative analysis of the mechanics of either the natural feature or his experiments. There are too many variables.

Although only the two thick beds were examined along the strike, several smaller beds were located in a 2-mile section across the strike. Since the term "bed" has been used, it would be reasonable to expect some sort of fold pattern to be traceable by connecting the beds into anticlines and synclines. There was no obvious arrangement of the known beds. Soil cover is always a problem in Piedmont geology. There may be other beds, as yet undiscovered, which are the missing limbs of anticlines and synclines.

A weak point in the argument lies in the dip changes observed in the large beds. In several places along their strikes the dips are reversed from west to east and from east to west. The explanation is probably more complex than simple folding alone. The orientation of the western bed is easily identified, since the eastern member is always the thinner. If both limbs of an anticline were exposed, this key identification would make it recognizable. The beds were not found in a reversed position at any point. This must mean that, granted that the graphite is in beds, only one limb of the fold pattern is exposed as the western bed, and the problem of the dip change remains unsolved.

The eastern bed is the widest, having an average thickness of 200 feet and a

maximum thickness of 360 feet. It is not a single, solid mass of schistose graphite but consists, rather, of graphitic schists interspersed with micaceous schist carrying no appreciable graphite. The graphitic zones average about 4 per cent graphite, although any minable width would have a much lower proportion because of the interspersed gangue material. This bed is, in general, a ridge-maker because the carbon is chemically inert and decomposes less rapidly than the more feldspathic gneisses in the region. Its low, rounded ridge is an easily marked feature, which helps make this band an exceptionally fine key horizon. The alternations of zones of graphite schist and mica schist of about equal width distinguish the eastern bed from any other.

A quarter of a mile west of the wide eastern bed there is a second characteristic key horizon. This western bed differs from the eastern in two major respects. The graphite content of the western bed varies from about 4 to a maximum of 35 per cent, whereas there is no known outcrop of the eastern bed which has more than 4 per cent graphite. In the western bed there are only two zones of graphite schist, a few feet apart. These two zones seem to be continuous. At every good outcrop of the bed, where the complete exposure is certain, both may be seen. Although neither half has a constant width, the western half is invariably the wider, averaging 40 feet in width, while the eastern half averages 19 feet. Neither varies more than a few feet from its average thickness. The mica schist separating them averages 10 feet in thickness. This tendency to maintain these uniform widths makes the western bed an unmistakable key horizon.

All the old mines and most of the prospect pits are located along the strike of the western bed. Early prospecting located all the graphite-rich zones. The old pits are still open enough to permit examination. Two old mining sites have the local names, "Old Lead Mine" and "Lead Mine Hill." The "wall-to-wall" graphite average of the former is 12 per cent; that of the latter is 35 per cent. Each of the old mining ventures failed because the ore could not be washed successfully and the quartz removed.

ORIGIN OF THE GRAPHITE

There are two possible origins for this graphite. One is magmatic and was proposed by Ebenezer Emmons[2] in 1856. He made a regional survey of the midland counties of North Carolina and noticed that the trend of the graphite was not parallel to what he considered to be the trend of metamorphics over the big area that he studied. His belief was that the graphite had come up across the foliation from below as magmatic carbon either by the reduction of CO_2 or from buried carbonates.

It has been pointed out that the strike of the foliation in the wall rock is parallel to the strike of the bands themselves as well as to the strike of the foliation of the schistose graphite. This observation is in direct opposition to that of Emmons. His conclusions must have been based on regional-reconnaissance methods rather than on strike mapping of the graphite itself. To have produced the observed structure by magmatic action, the graphite would have had to be injected along parallel planes of weakness followed by regional metamorphism. This metamorphism would have had to be so directed that the foliation developed in the country rock and graphite was parallel to the strike of the bands. Supporting evidence

[2] *Ibid.*, p. 222.

in the form of offshoots and cross-shoots, vein banding, or obvious crystalline graphite could not be found. However, it is the extreme regularity of the western bed with its two member-beds that is the major obstacle to this magmatic hypothesis. It is difficult to conceive of a set of weakness planes which would remain as separate as these two halves of the western bed over so great a distance as 10 miles. The two parts are separated by a few feet of schist, in which there is very little graphite, and that in the form of disconnected specks rather than stringers. The members themselves are distinguishable because the western one is invariably the wider. Since the case for magmatic carbon seems weak, we must consider an alternative that this graphite was the product of metamorphism acting on a carbonaceous sediment.

In seeking the mode of origin from the field evidence, it seems that the sediments must have been quite regular. The stratum which contained the carbon was probably underlain and overlain by sediments quite free from carbon. The estimate of regularity is based on the even widths of the members of the western bed. The consistent thicknesses probably indicate marine, rather than continental, sediments. The mica and quartz content of the graphite beds show that these beds were probably carbonaceous, arenaceous shales before metamorphism. Since the enclosing schists do not seem to carry any appreciable quantity of graphite, it is quite logical to suppose that these were practically free of carbon.

The beds are now found in what appears to be a folded position. The foliation of the graphite schist is parallel to the strike of the graphite beds. In order to obtain this structure it would have been necessary for the original folding which deformed the beds to continue on with the same orientation of the compressive forces until the dynamic metamorphism was complete. This is a special case and an assumption which is admittedly a weak point in the explanation of the origin. Rocks of this highly deformed nature would be expected to contain beds that marked out pitching anticlines and synclines rather than the relatively straight strike pattern which was observed. Until the entire area has been mapped and the origin of the nongraphitic gneisses and schists definitely determined, this assumption is the simplest explanation of the straight strike pattern. The dynamic metamorphic history was completed when the regular, nonplunging folds were formed and the graphite had been made from the sedimentary carbon.

Each member of each bed is believed to have been a depositional unit. Variations in the graphite content of the beds along their strike could be due to depositional irregularities. Quartz is the major mineral in all the beds. It seems to be of three different ages and to have three entirely different origins. Thin sections of these graphite schists show that some of the quartz grains are strained. These were in all probability present during the dynamic metamorphic stage and owed their deformation to this stress. The unstrained quartz occurs in two ways. Most of it is disseminated throughout the entire schist. This portion of the quartz is believed to be due to a complete *lit-par-lit* injection of much of the Raleigh area. The details and magnitude of this injection demand a more thorough investigation of the quartz of this age than can be made in the injected graphite alone. The author has never seen schist or gneiss in the Raleigh area that was not shot full of this disseminated quartz. Its injection is not comparable to the injection banding

often seen in a gneiss. There is no tendency for the quartz to be localized in bands. The rock has the appearance of having been literally stewed in quartz juice so that the quartz dispersion is homogeneous. Even those schists which carry a large percentage of graphite have a tremendous amount of quartz.

The third type of quartz occurs in stringers, small pegmatites, and veins up to 18 inches thick. This quartz is younger than any other mineralization, as the veins cut all the other structures and fill many of the joint cracks. It is believed that this third type of quartz is associated with the intrusion of granite east of Raleigh and is late Paleozoic in age.

The criteria for separating these three types of quartz are: strain for the oldest, ubiquity and lack of strain for the injected quartz, and vein structure for the youngest.

Kyanite and garnet are also found in a few places. Most of the garnet occurs in a dense veinlike zone about 6 inches wide at the Lead Mine Hill site in the western bed. The garnets preserve their dodecahedral form but are clustered closely together. The kyanite occurs in typical bladelike crystals, which definitely cut the schistosity. There is no tendency for these crystals to be aligned along the foliation. Other garnets occurring individually show that the foliation may be cut by the dodecahedron or may bend around the crystal. These garnet occurrences are believed to be contemporaneous with the metamorphism. However, the kyanite is definitely postdynamic metamorphism. The specific relationship of the kyanite age to the quartz ages has not been determined.

CONCLUSIONS

1. It is believed that these graphite beds were once carbonaceous sediments that have been folded and metamorphosed to their present condition.

2. Since graphite is chemically inert, it has not been affected by hydrothermal alteration, and the original bed structure is still recognizable. This fact will be of vital importance in deciphering the true structure of the highly metamorphosed schists and gneisses of Wake County. These rocks are so differentially altered that they cannot now be mapped as depositional units. The Raleigh graphite furnishes easily distinguishable key horizons.

ACKNOWLEDGMENT.—The author is indebted to the faculties of North Carolina State College and the University of North Carolina for assistance in interpreting these features.

DIABASE DIKES OF THE FRANKLIN FURNACE NEW JERSEY, QUADRANGLE[1]

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United States Geological Survey

ABSTRACT

Four of the numerous dikes mapped on the areal geology sheet of the *Franklin Furnace Folio*[2] cannot be included among those therein described as "mostly basic, including nepheline tinguaite, leucite tinguaite, and camptonite" of "post-Ordovician" age. Field and petrographic study and chemical analyses show that these four dikes form a closely related group, sharply distinguishable from all other known dike rocks of the area. They are in all essential respects similar to the Triassic diabases of other parts of New Jersey and the Atlantic seaboard states, and there is no reason to question their Triassic age.

LOCATION OF DIKES

The locations of the four dikes are as follows:

1. East of Howell Pond, near the locality marked "Pinckneyville," in the southwestern part of the quadrangle. This occurrence will be called the "Pinckneyville dike" in the following discussion. On the rectangular co-ordinate grid system of the Sussex County, N.J., *Base Map* (N.J. Highway Dept.–Federal Works Agency, Public Roads Administration, 1939), the dike extends for about $\frac{1}{4}$ mile N. 45° E., approximately diagonally across the square whose co-ordinates are 1991–92 and 804–5. The length of the dike is at least twice that indicated on the *Folio* map. The dike forms the topographic ridge and probably extends somewhat farther to the east than is indicated on that map.

2. About 2 miles east-northeast of the Pinckyneyville dike, in the vicinity of Sussex Mills. Here this is called the "Sussex Mills dike." It extends in a northeasterly direction for about $\frac{1}{8}$ mile, as indicated on the *Folio* map, and is about 30 feet wide. Its co-ordinates are 2000–2001, 805–6.

3. Near the middle of the southern edge of the quadrangle, about a mile south of the old Mount Sherman prospect. This is called the "Mount Sherman dike," and its co-ordinates are 2011–12, 793–94. It is the smallest of the four dikes, being only some 10 feet long and a few feet wide. Its strike, approximately N. 50° W., is different from that of the others as shown on the *Folio* map. It is about 200 feet east of a farmhouse.

4. Near the hamlet of Russia, in the southeast corner of the quadrangle. Its co-ordinates are approximately 2035–36 and 804–5. It extends for some 200 feet, about N. 80° E., along both sides of the road. It has been freshly blasted along the roadside and offers good exposures.

OTHER DIKES

All the other known post-Ordovician dikes of the quadrangle intrude either pre-Cambrian marble (Franklin limestone) or Cambrian and Ordovician rocks, whereas the diabase dikes cut pre-Cambrian gneiss—the Pinckneyville dike being in the Pochuck gabbro gneiss, while the Sussex Mills, Mount Sherman, and Russia dikes are in the Losee diorite gneiss. All four diabase dikes occur in the southern third of the quadrangle, remote from the other known dike rocks of the northern half of the quadrangle. Whether

[1] Published by permission of the director, United States Geological Survey.

[2] *U.S. Geol. Surv. Geol. Atlas, Franklin Furnace Folio 161* (1908).

similar diabasic intrusives occur to the south, in the Lake Hopatcong quadrangle, is not known; but at least through 1944 no Triassic rocks, igneous or sedimentary, had been recorded in either Sussex or Warren County.[3] Near the extreme southern border of Morris County, however, where Morris, Somerset, and Hunterdon counties adjoin, an isolated diabase dike is indicated on the geological map of New Jersey (J. Volney Lewis and Henry B. Kümmel, 1910–12). It is about 6 miles north of the belt of Triassic sedimentary and igneous rocks crossing the state and fully 15 miles from the Franklin Furnace quadrangle. It is mapped as intruding Byram gneiss and trends northeast-southwest (conforming with the regional structure). It is about $\frac{1}{3}$ mile in length.

A careful examination has been made of all mapped dikes in the Franklin Furnace quadrangle, as well as of a great number that have not been mapped, and none of them, except the four herein described, is diabase. No Triassic sedimentary rocks are known in the region, and the wide Triassic belt of sedimentary and igneous rocks crossing New Jersey is 20 or more miles to the southeast. Nevertheless, the occurrence of these four dikes is believed to be of significance in that it indicates that the locus of the alkalic Beemerville nepheline syenite and associated volcanics is at no great distance from the outliers of the geotectonic region of Triassic faulting and vulcanism. In a paper now in preparation, other instances of association of nepheline syenite and known Triassic igneous rocks will be discussed.

In a third paper four other bodies of intrusive rock forming a closely related group, markedly different from both the Beemerville alkalic suite and the presumably Triassic diabases herein described, and designated the "Franklin Furnace Kugel-minettes," will be discussed. On the sketch map (Fig. 1) are shown the geographical relations of the diabase dikes, the Beemerville alkalic complex, the Kugel-minettes, and the northern boundary of the New Jersey Triassic sedimentary-igneous belt, which is indicated by the heavy dashed line.

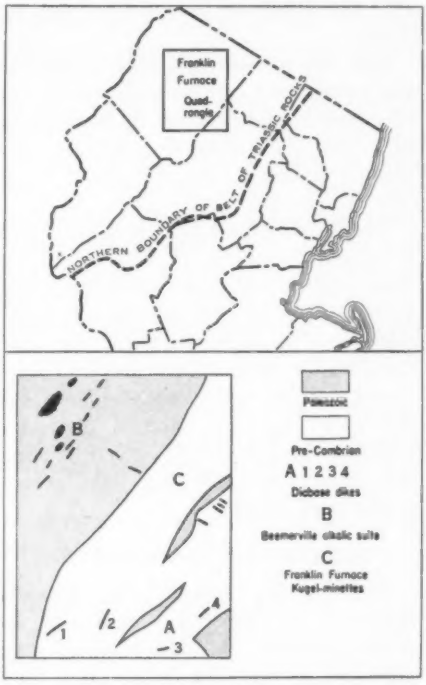

FIG. 1.—Map showing diabase dikes in the Franklin Furnace quadrangle.

FIELD RELATIONS OF THE DIKES

The Pinckneyville dike is most easily reached from the road east of Howell's Pond. It forms massive ledges, and its contacts with the intruded Pochuck gabbro gneiss show no observable contact-metamorphism.

[3] Personal communication from Meredith E. Johnson, state geologist of New Jersey, December 14, 1944.

The Sussex Mills dike is similar in topographic expression to the Pinckneyville dike; it, too, forms a low ridge which, with a few gaps, can be followed for many hundred feet; in a few places contacts with the gneiss can be seen. The diabase varies irregularly from fine grained to rather coarse grained. Probably the difference is due to a chilling effect.

The Mount Sherman dike is without any topographic expression and is noticeable only by its contrast in appearance with the light-colored gneiss, across whose foliation it cuts sharply. It has a dull, stony aspect.

The Russia dike differs from the others in several respects. It has undergone some folding, and superficially resembles a greenstone and, without detailed study, might be thought part of the pre-Cambrian metamorphic complex. Contorted veinlets with much epidote cut the rock, and the dike is distinctly schistose, and slickensiding can be seen. No contacts with the local light-colored gneiss were found.

PETROGRAPHY OF THE DIABASES

The diabasic rocks of the Atlantic seaboard are generally known to be fairly uniform in mineralogical and chemical composition, and these four occurrences are no exceptions to the rule.

The Pinckneyville dike is a fine-grained, dark-gray rock, of rather uniform texture throughout its exposure, showing little or no weathering. Grains of pyrite, 1–2 mm. across, are perfectly fresh and show no signs of oxidation.

The effect of rapid cooling, on contact of the dike with the enclosing granite gneiss, is evident on microscopic study. The slowly cooled rock is holocrystalline, with excellent development of all the major minerals, whereas, in the chilled border phases, an obscurely crystallized groundmass contains isolated well-developed early-formed crystals. Intermediate plagioclase, diopsidic pyroxene, quartz, and black ore grains comprise the bulk of the well-crystallized diabase. The quartz usually occurs in isolated rounded areas, often with pale-green chlorite. The ore grains commonly show a characteristic skeletal development. In very minor quantity are found sphene, always of opaque grains, and apatite. In contrast, the chilled border phase shows no quartz or accessory minerals; only a crowded semi-glassy brown groundmass in which euhedral plagioclase crystals and diopsidic pyroxene, in single crystals and complexly intergrown composite groups, are strewn.

A thin section across the contact of the diabase and the granite gneiss shows a hair-line contact, with no suggestion of any contamination of the diabase by the adjoining gneiss. The gneiss, which consists essentially of sericitized microcline and quartz, likewise shows no indication of mechanical deformation by the intrusive diabase.

The Sussex Mills dike consists of two main varieties, which, however, grade into each other, one of distinctly fine grain, the other of coarser. There is also a variety in which hydrothermal alteration processes have caused marked changes, producing a rock rich in quartz and epidote, similar to the Russia rock described below.

The fine-grained Sussex Mills rock is dense, with tiny feldspar cleavage surfaces and minute pyrite grains breaking up the dull appearance of the fresh rock surface. Weathering is almost absent. The other variety is distinctly coarser, with a granularity suggesting a gabbro. Pyrite and cleavage surfaces of feldspar and of pyroxene grains are visible. The microscope shows, however, that the

mineralogy of the two rocks is very similar. The analysis given below is of the fine-grained rock.

The finer-grained variety of the Sussex Mills rock shows, under the microscope, prophyritic crystals of perfectly fresh calcic plagioclase and pyroxene, in an ill-defined yet holocrystalline groundmass consisting of these two minerals, with subordinate amounts of ilmenite-magnetite, leucoxene, chlorite, biotite, and pyrite. The groundmass is characterized by a rectangular network of ilmenite-magnetite crystallites, which are not present in the coarser variety.

The coarser-grained variety of the Sussex Mills rock consists mainly of pyroxene and calcic plagioclase, with subordinate quartz and sodic plagioclase (micropegamatite), also apatite, ilmenite-magnetite, leucoxene-sphene, and indefinite chloritic areas. With these last are some wisps of green hornblende or brown biotite. Microscopic fractures, filled with green chlorite, traverse the section. The calcic plagioclases show a slight turbidity due to sericitization.

The difference between the two varieties is presumably due to different conditions of crystallization; the fine-grained rock being chilled after the development of the porphyritic plagioclases and pyroxenes and the magnetite-ilmenite grid structure. In the coarser-grained variety there was a relatively slow cooling, sufficient to permit a uniform crystallization, and agglomeration of the grid-material into equant grains.

The hydrothermally altered variety of the Sussex Mills dike is a dull greenish-gray stony rock, seamed with green bands of epidote-quartz, which project from the weathered surface. Slickensides are preserved on these harder bands. Microscopically, the typical diabase texture has been obliterated; no pyroxenes remain and but little of the calcic plagioclase. Most of the feldspar present appears to be clear, not turbid, is untwinned, and near oligoclase-andesine in composition, judging by the refraction. Clear quartz, with much yellowish epidote, is abundant. Some of the epidote enclosed in quartz is euhedral. Green chlorite is abundant in the quartz-epidote areas. Leucoxenized titanited grains are strewn about the section, and there is some calcite.

This rock shows clearly the relation between the very similar Russia rock and the other diabases. Without it there might be some doubt as to whether the Russia rock should be considered a diabase. Finding this altered diabase also raises the question as to whether certain other rocks of diabasic character, referred on the basis of alteration to an age older than Triassic, may not actually be Triassic.[4]

The Mount Sherman dike, though only 2 feet wide, nevertheless shows a contrast between the middle coarser part and the finer-grained chilled borders. An interesting feature is the presence of numerous masses of bright-green serpentine or chlorite, whose sharp outlines strongly suggest replacement of olivine. The reddish proxene is, in general, completely unaltered, but a few grains contain a core of the green mineral. These replacements are present in both the coarse- and the fine-grained rock. In the latter there is a distinct porphyritic development, mostly of plagioclase, with less pyroxene; the groundmass is an obscure aggregate of extremely fine grain, probably, in part, glass. Microscopic veinlets that appear to be of the same substance that has replaced the (possible) olivine also traverse the rock. No quartz was seen.

[4] Anna I. Jonas, "Pre-Cambrian and Triassic Diabase in Eastern Pennsylvania," *Bull. Amer. Mus. Nat. Hist.*, Vol. XXXVII (1917), pp. 173–81.

The Russia dike is uniform, of fine grain, gray-green in color, and is fresh and tough. Thin veinlets of quartz-epidote cut the dike, which is somewhat schistose and slickensided. The contact with the enclosing light-colored gneiss was not observed. Except for a few specks of pyrite, no minerals can be distinguished in the hand specimen. Microscopically, it resembles the coarse variety of the Sussex Mills dike, except that little or no quartz is present. However, the Russia dike shows more alteration effects than any of the others. There are no porphyritic plagioclase crystals; the pyroxene, which is distinctly red, shows peripheral replacement by wispy amphibole; and much of the coarse ilmenite has gone over to leucoxene.

These four analyses are very similar and bear a close resemblance to published analyses of Triassic diabases from other localities. In the C.I.P.W. classification, the norm of the Pinckneyville, Sussex Mills, and Russia diabase dikes is III.5.3.4; the norm of the Mount Sherman differs but slightly, being III.5.4.4; this difference reflects a slightly higher ratio of CaO to alkalies.

The norms of the Pinckneyville and Sussex Mills rocks are very similar, as the petrographic description of these would indicate. The Mount Sherman rock shows the highest An–Ab ratio, presumably correlated with the small size of the dike. Rapid chilling would produce a fine-grained compact rock, which might be less susceptible to albitization or similar deuteric effects. The Russia rock, on the other hand, has the lowest An–Ab ratio, and it is this dike (with the altered phase of the Sussex Mills dike) which shows the most effect of hydrothermal activity, such as destruction of the early-formed calcic-plagioclase crystals, formation of quartz-epidote veinlets, and formation of leuxocene and pyrite.

TABLE 1*

CHEMICAL ANALYSES OF THE DIABASE DIKES FROM THE FRANKLIN FURNACE QUADRANGLE

(Charles Milton, Analyst)

	A	B	C	D
SiO_2	49.25	49.49	49.03	47.42
Al_2O_3	13.62	14.63	13.72	13.99
Fe_2O_3	4.22	5.39	4.11	1.97
FeO	9.18	8.83	10.44	11.49
MgO	4.43	4.62	5.57	6.62
CaO	7.63	8.03	9.96	7.81
Na_2O	2.61	2.70	1.90	3.30
K_2O	2.00	1.60	0.81	1.54
H_2O^+	2.16	1.56	1.89	2.71
H_2O^-	0.07	0.10	0.06	0.11
CO_2	.....	.....	.....	trace
TiO_2	3.31	2.56	2.43	2.30
P_2O_5	1.23	0.46	0.08	0.30
S	0.18	0.23	0.09	0.26
MnO	0.26	0.20	0.23	0.15
BaO	0.02	0.03	0.02	0.00
ZrO_2	.....	.....	.....	0.05
Total	100.17	100.43	100.34	100.02
Density	2.92	3.01	3.00	3.06

*A, diabase, Pinckneyville, Andover Township, Sussex County, New Jersey; B, diabase, Sussex Mills, Sparta Township, Sussex County, New Jersey; C, diabase, Mount Sherman, Sparta Township, Sussex County, New Jersey; and D, diabase, Russia, Jefferson Township, Morris County, New Jersey.

TABLE 2

NORMS OF THE DIABASE DIKES FROM THE FRANKLIN FURNACE QUADRANGLE

	Norms			
	Pinckneyville (III.5.3.4)	Sussex Mills (III.5.3.4)	Mount Sherman (III.5.4.4)	Russia (III.5.3.4)
Quartz	5.6	4.1	4.9	.....
Orthoclase	11.7	9.5	5.0	8.9
Albite	22.0	23.1	15.7	27.8
Anorthite	19.5	22.8	26.7	18.9
Diopside	8.2	11.7	18.7	15.8
Hypersthene	14.8	13.2	16.3	.....
Olivine	.....	.....	.....	17.3
Magnetite	6.0	7.9	6.0	2.8
Ilmenite	6.1	4.9	4.6	4.4
Water, etc.	5.2	3.0	2.3	3.9
Total	99.1	100.2	100.2	99.8

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REVIEWS

Dating the Past: An Introduction to Geochronology. By FREDERICK E. ZEUNER. London: Methuen & Co., Ltd., 1946. Pp. xviii+444; figs. 101; pls. 24. 30*s*.

Dating the Past is a companion volume to Zeuner's *The Pleistocene Period*. While the latter deals mainly with the evidence for the relative and absolute chronology of the Pleistocene, the present book is concerned with the various methods of dating the past, with geochronology in general, and with its application to archeology in particular (p. v). Geochronology is defined as "the science of dating in terms of years those periods of the past to which the human historical calendar does not apply" (p. 5). The word was probably first employed in this sense by Gerard De Geer in 1912 with reference to his measuring time by means of the late-glacial and postglacial varved clays.[1] Knowledge of age and time is necessary for perspective and understanding, as is reflected in the layman's ever recurring questions: How old? How long ago? How long a time did it take?

Zeuner treats four geochronological means (p. 5) and divides the treatise into four corresponding parts, viz.: (1) tree-rings: dating early history and late prehistory, especially in North America (pp. 6–19); (2) varved clays: dating the Metal ages, New and Middle Stone ages, and the climatic phases which followed the Ice Age (pp. 20–109); (3) the varying distribution of the solar radiation: dating the Old Stone Age, the phases of the Ice Age and the Pluvial phases of the warmer countries (pp. 110–304); and (4) radioactive disintegration: dating the history of the earth and life before the arrival of man (pp. 305–85).

The expressions "tree-ring analysis," which has been employed by others, and "varved clay analysis" and "varve analysis," which are probably new (pp. 5, 6, 20), seem improper since for use in dating and climatic research tree-rings and varves are not resolved into their constituent parts. "Tree-ring research," "tree-ring science," "varve research," and "varve science," or "varvology," seem more appropriate.

Tree-rings have long been studied from many angles. Their chronological importance is based on the fact that their width is notably influenced near the dry limit of tree growth by the water supply available during the growing season and near the cold limit by the temperature of the season of growth. The modern ring research has two distinct aims: (1) dating of prehistoric ruins (by means of wooden beams), and (2) investigation of past variations in climate, especially in rainfall. The study as a whole is usually called "dendrochronology," and "tree-ring hydrology" and "dendroclimatology" are taken as subdivisions. The first method of using tree-rings in archeological dating and the most-used method in climatic studies was devised by A. E. Douglass of the University of Arizona. This method is outlined with reference to Waldo Glock's detailed description. Some examples of ruin-dating in the American Southwest are given, and cycles in radial growth, especially the eleven-year, sun-spot cycle, are discussed (pp. 12–19). It should be added that Harold S. Gladwin has developed another method of tree-ring correlation for beam-dating[2] and also that Edmund Schulman of the University of Arizona has carried out important dendroclimatic research.

Varve is the distinguishable annual deposit of any sediment. It is the equivalent to the annual ring in wood. Varves occur in both organic and inorganic sediments. True or supposed varves in various pre-Pleistocene sediments are listed on pages 36–38. However, only three kinds are of chronological importance so far, viz., the glacial varved clay, the postglacial varved silt of northern Sweden, and the varved gyttja in Faulensee, Switzerland, too recently described to be considered by Zeuner.[3]

[1] "A Geochronology of the Last 12,000 Years," *Compt. rend. XI^e Cong. géol. internat.* (Stockholm, 1912), pp. 241–53.

[2] "Tree-Ring Analysis: Methods of Correlation," *Medallion Papers No. XXVIII* (1940), Gila Pueblo, Globe, Ariz. Reviewed in *Amer. Anthropologist*, Vol. XLVIII (1946), pp. 433–36.

[3] Max Welten, "Pollenanalytische, stratigraphische und geochronologische Untersuchungen aus

The study of varves was originated by De Geer in 1878, and practically all phases of the method in use were developed by him.

The postglacial varved silt is still being formed, but since its latest beds are below the level of the Baltic, the count, which comprises 7,760 varves, does not extend up to the present. However, by means of a careful chronological study of the land rise, Ragnar Lidén found that the last measured varve was deposited about A.D. 920; and the first postglacial varve was thus laid down about the year 6840 B.C. (p. 26).

The postglacial varved clay is conformably underlain by glacial varved clay, which has enabled a direct backward extension of the varve chronology. This glacial chronology embraces the wide belt of ice release from northeastern Scania (56° N.), the southernmost province of Sweden, to Stugun in northern Sweden (63°10′ N., 15½° E.). In the opinion of Sigurd Hansen and of the reviewer it cannot be extended over southwestern Scania and the Danish Islands (cf. pp. 27, 29, 31). Not all data have been published, but this glacial-varve chronology probably comprises 4,500 to 5,000 years. Consequently, the last ice border stood in northeastern Scania some 13,500 years ago; and 13,500 years is the length of the combined postglacial and late-glacial Swedish varve chronology.

In Sweden the bisection of the ice remnant at Stugun and the accompanying change from glacial to postglacial varved clay is, after De Geer's suggestion, taken as the demarcation line between the Glacial and the Postglacial. So defined, the Postglacial began in 6840 B.C. and comprises the last 8,800 years (p. 28). Other suggested demarcation lines—the beginning of ice retreat from the Fennoscandian moraines (p. 28), the attainment of the modern temperature level in the southern and greater parts of the glaciated regions,[4] and the arrival of the modern forest in Denmark[5]—are based on recognizable climatic changes for long-distance correlation and broader application. They place the transition at 9000–10,000 B.P. (before the present).

Matti Sauramo's varve studies in Finland and mine in North America are summarized (pp. 32–35). Since two figures and statements attributed to me are incorrect, my chronology[6] of the ice retreat from Mattawa, Ontario, to James Bay is here restated: 1,000 years estimated + 2,000 counted of retreat + 2,000 in part counted of oscillations in the Cochrane region + 1,000 years estimated for final retreat to James Bay: total 6,000 years. The subsequent postglacial age is, as just stated, taken to be 9,000 years, and the ice border at Mattawa is so given the age of 15,000 years. The important Mattawa ice border can also be dated thus: The Cochrane ice-border oscillations and moraines are probably correlatives of the Fennoscandian moraines, now dated by Sauramo at about 8900–8000 B.C., or roughly 11,000–10,000 B.P. If the drainage of Lake Ojibway-Barlow, which just preceded the first ice readvance to and beyond Cochrane, therefore be dated at 11,000 B.P., and if 2,000 counted and 1,000 estimated years be added, the date of the border at Mattawa will be 14,000 B.P.

The application of the varved-clay chronology to archeological dating is usually by a round-about route (p. 46). Some stages and shorelines of the Baltic have been directly geochronologically dated, say within one hundred years. And by means of pollen grains in varve series the regular appearance, or the maximum frequency, of characteristic forest trees as well as some distinct climatic ages have been rather exactly dated. For instance, the influx across central Sweden of salt water into the Baltic basin which introduced the Yoldia stage took place from 8000 to 7900 B.C., the transition to the Ancylus stage about 6500 B.C., and that to the Littorina stage about 5000 B.C. (pp. 51–53). The alder became a regular constituent of the forest about 6350 B.C., and the spruce between 1000 and 950 B.C.; and the birch reached it greatest frequency about 4200 B.C. The temperature was (on 63°10′ N.) highest during the age 5000–3500 B.C., and it dropped distinctly around 1000 B.C.[7]

With the help of such starting-points other Baltic shorelines and stages and other pollen

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dem Faulenseemoos bei Spiez," *Veröff. Geobotan. Inst. Rübel in Zürich*, No. 21 (1944). Reviewed in *Amer. Jour. Sci.*, Vol. CCXLIV (June, 1946), pp. 442–47.

[4] Ernst Antevs, "Late-glacial Correlations and Ice Recession in Manitoba," *Geol. Surv. Canada, Mem. 168* (1931), pp. 2, 6.

[5] Johs Iversen, "Plantevaekst, Dyreliv og Klima i det senglaciale Danmark," *Geol. Fören. i Stockholm Förhandl.*, Vol. LXIX (1947), p. 68.

[6] Pp. 18, 19, 29, 33, 34 of ftn. 4.

[7] Erik Fromm, "Geochronologisch datierte Pollendiagramme und Diatoméenanalysen aus Ångermanland," *Geol. Fören. i Stockholm Förhandl.*, Vol. LX (1938), pp. 365–81; see pp. 379–80.

horizons and pollen profiles have been dated approximately and so have archeological finds associated with shores or located in peat bogs. In the last instance the possibility of dating is usually dependent on climatic changes enabling correlation.

The history of the Baltic during the last 11,000 years is outlined mainly after Sauramo's summaries (pp. 47–55). The pollen analysis and other botanical ways of studying the changes of the forest composition and thus of the climate are described as well as the deduced climatic stages (pp. 56–72). A long chapter reviews important archeological sites from the past 15,000 years in northwestern Europe (pp. 72–109). The use of the floral and climatic succession as basis for chronology beyond Fennoscandia meets two well-known complications, viz., the distinct zoning and the northward migration of the zones as the ice border withdrew and the varying rate of spread of the species (p. 104). However, dated, distinct, and widespread climatic changes and geological events enable at least many approximate correlations and datings (p. 108).

Part III, "Dating the Old Stone Age, the Phases of the Ice Age and the Pluvial Phases of the Warmer Countries," begins with a chapter on the relative and absolute chronology of the Pleistocene (pp. 110–45), which is a summary of Zeuner's book *The Pleistocene Period*, reviewed in *Journal of Geology*, Volume LV (1947), pages 446–50. The absolute astronomic chronology of the glacials and interglacials shows a correspondence with the relative, stratigraphic-climatic chronology which is difficult to explain as mere chance. However, "if anybody regards the application of the absolute time-scale as unconvincing, it is open to him to discard this part of the Pleistocene chronology altogether. That, however, will not effect the value of the relative, climatic, chronology" (p. 110). While the astronomic chronology might really apply, still, as was pointed out in the above-mentioned review, not every pronounced minimum of summer insolation, especially not the first in the group of two or three, need have been represented by a continental glaciation; and, as a consequence, some of Zeuner's dates perhaps need adjustments. For instance, ice sheets were probably not formed upon the radiation minimum 115,000 B.P., that is, there was probably no Würm, or LG1, continental glaciation, but the Warthe (and the Iowan) may, rather, represent the radiation minimum of 72,000 B.P.

The dating of the Paleolithic finds and sites in temperate Europe is made on geologic and/or paleontologic grounds and thus fitted into the astronomic chronology. It is not made with implements used as characteristic fossils (p. 146). Industries dated in this manner have been used in building up relative and absolute chronologies of the Paleolithic of central Europe, France, and Britain (p. 200). If, as suggested, the first ice sheets of the Last Glacial were developed upon the radiation minimum of 72,000 B.P., the Mousterian culture of the Neanderthal man which belongs to the Last Interglacial and the next stage of extensive glaciation would be considerably extended upward in the time scale to, say, 50,000 or 40,000 B.P., which would be in accord with archeology.[8]

Other chapters deal with the "Chronology of the Pleistocene and Palaeolithic of the Mediterranean Area" (pp. 201–43), "Climatic Phases, Early Man and Human Industries in Africa, Asia, Australia, and America" (pp. 423–81), and "Chronology of Early Man and His Cultures" (pp. 281–304). The *Pithecanthropus* group of *Homo* is believed to have lived prior to 400,000 B.P. (p. 295), the Neanderthal group during the Last Interglacial and the first continental glaciation of the Last Glacial (p. 296), and the *Homo sapiens* group may have existed at least during the last quarter-million years (p. 298). The Neanderthal man produced the Mousterian flake industry and *Homo sapiens* the blade cultures and perhaps the hand-ax industry (p. 302).

Part IV, "Dating the History of the Earth and of Life before the Arrival of Man," begins with a chapter on the measurement of geological time previous to the Pleistocene Ice Age (pp. 305–33). It reviews the methods of estimating and measuring time based on organic evolution, sedimentation, accumulation of salts in the oceans, cooling of the earth, and radioactive decomposition, which last method has superseded the others. The oldest radioactive minerals are about 1,750,000,000 years old, and Arthur Holmes has estimated the earth's minimum age at 1,900,000,000–2,000,000,000 years. Of this about three-fourths belong to the Pre-Cambrian, the Cambrian beginning some 500,000,000 years ago (pp. 333, 335, 337). The dates serve as bases for estimates of the time rates of

[8] Robert J. Braidwood, "Geochronology, and Geological Factors Relative to Man's Past," pp. 43–56 in *Human Origins: Selected Readings* (2d ser.; Chicago: University of Chicago, 1946). See p. 52.

weathering, erosion, mountain-building, and other geological processes (pp. 337–49).

The last chapter, "Biological Evolution and Time," stresses the significance of geochronology for studies on the evolution of life (pp. 350–85). In a table the minimum time of existence of several animal groups is given, which for man is 1,000,000 years, for manlike apes 50,000,000, and for mammals 160,000,000 years (p. 351). The organic evolution has proceeded at a varying pace. It has been fast and has produced many new types within relatively short spaces of time (or perhaps 50,000,000 years), then slow during longer ages (pp. 352–62). While subspecies have developed within the last 7,500 years (pp. 364–66), full species of terrestrial animals ranging from mammals to insects have required at least 500,000 years and species of aquatic animals still longer time (pp. 375, 378). "With the establishment of geochronological time-scales a new element has entered into the study of evolution and . . . . , as these time-scales are improved, they are bound to become increasingly valuable as measures of the actual rates of evolution" (p. 384).

This is a very impressive book, and the seven years spent in its preparation have clearly been years of intensive research and deliberation. It should serve as a useful handbook and as an inspiration to students of several sciences dealing with the past.

ERNST ANTEVS

Glacial Geology and the Pleistocene Epoch. By RICHARD FOSTER FLINT. New York: John Wiley & Sons, Inc.; London: Chapman & Hall, Ltd., 1947. Pp. 589; pls. 6; figs. 88. $6.00.

The appearance of this book by Professor Flint of Yale University marks a milestone in the history of Pleistocene studies in that it is the first formal text by an American author to cover the fields of glacial and Pleistocene geology. In addition, it provides a comprehensive summary of a vast literature and a means of evaluating problems on which further research is needed. So many fundamental problems are discussed that a full review will not be attempted.

About one-third of the text concerns glacial processes, land forms, and deposits, while most of the remainder is devoted to Pleistocene history. There are six folded plates which show: existing glaciers, storm tracks, pack ice, and selected ocean currents in (1) the northern hemisphere and (2) the southern hemisphere; (3) inferred glaciers, pack ice, and storm tracks in the northern hemisphere during the maximum of Wisconsin glaciation; (4) distribution of loess and Urströmtäler in Europe; (5) glaciated areas in western United States; and (6) glaciated areas of Europe. Numerous additional glacial maps occur as figures in the text. The text figures include seventy maps and diagrams but only eighteen photographs. About nine hundred references, a high percentage of them cited in the text, are listed separately. There is full coverage of important foreign titles.

In an introductory chapter the historical background is briefly summarized. This is followed by nine chapters on glacial geology dealing with the following topics: glacier ice and glacier motion, regimen of glaciers, existing glaciers, glacial erosion, glacial deposits and their interpretation, glacial drainage, ice-thrust features, and eolian deposits. The chapter on existing glaciers covers the entire world and is analytical rather than descriptive. In the treatment of glacial deposition the close relations between till and stratified deposits are emphasized. T. C. Chamberlin's terms, "dumping," "pushing," and "lodgement," are revived for the three principal processes in till deposition, and distinction is made between "basal till" formed by lodgement and "superglacial till formed by ablation.

The subject of Pleistocene history is introduced by a chapter on glacial stratigraphy in which full consideration is given to criteria for the differentiation of drift sheets, the problem of the Pliocene-Pleistocene boundary, and Pleistocene nomenclature. It is suggested that the terms "Tertiary" and "Quaternary" be dropped and that "recent" and "postglacial" be used only in an informal sense. The Pleistocene is considered an epoch and is provisionally defined on the basis of glacial climates and fossil vertebrates to include all of post-Pliocene time. The classification for the North American Pleistocene is given in Table 1.

In the next six chapters there follows a concise review of the Pleistocene record throughout the world. This section is more than an up-to-date regional summary in that it includes significant original interpretations by the writer. Comparisons between foreign glacial features and similar features in North America are especially helpful. The final chapter in the section considers methods of estimating Pleistocene time.

The concluding chapters in the book deal with: changes in level of land and sea, glacial

and interglacial climates, postglacial climates, causes of glacial climates, and the Pleistocene fossil record. The causes of glacial climates are narrowed to three groups of hypotheses based on: (1) topographic changes, (2) periodic planetary changes, and (3) variations in solar heat. A "solar-topographic hypothesis," offered by the writer, is a modification and extension of older topographic hypotheses and Simpson's solar hypothesis. The Milankovitch hypothesis, advocated by F. E. Zeuner in two recent books on the Pleistocene (*The Pleistocene Period: Its Climate, Chronology and Faunal Successions*

TABLE 1

Era	Period (System)	Epoch (Series)	Age (Stage)	Sub-age (Substage)
Ceno-zoic	Ceno-zoic	Pleisto-cene	Wisconsin	Mankato Cary Tazewell Iowan
			Sangamon Illinoian *Yarmouth* Kansan *Aftonian* Nebraskan	
		Plio-cene		

[1945], and *Dating the Past: An Introduction to Geochronology* [1946], is opposed on seven different counts.

In looking back over the contents of the book one is impressed by the years of compilation and careful consideration along with constant revisions that must have preceded the final copy. All students of the Pleistocene, and others in allied fields, welcome it as an invaluable text and reference book.

LELAND HORBERG

Geological Expedition to the Lesser Sunda Islands under the Leadership of H. A. Brouwer. Vols. II, III, and IV. Amsterdam: Noord-Hollandische Uitgevers Maatschappij, 1940–42. 17.00, 17.00, and 22.50 gulden.

The first volume of this four-volume series was well reviewed in considerable detail by Carey Croneis in 1940 (*Journal of Geology*, Vol. XLVIII, pp. 553–56). Attention should now be called to the three succeeding volumes whose appearance during the war has completed the set of formal reports prepared by the members of this important expedition under the leadership of Dr. Brouwer. They extend and round out the studies whose partial presentation in Volume I in 1940 aroused so much interest. Dr. Brouwer and his colleagues should receive the belated appreciation of all geologists interested in this portion of the globe which reveals so much that bears vitally on various diverse geologic problems. Timor, in particular, because of its exceptional features, has aroused hopes in not a few geologists that they may sometime have an opportunity to visit it.

Volume II, of 395 pages, presents the following contributions: H. A. Brouwer, "Geological and Petrological Investigations on Alkali and Calc-Alkali Rocks of the Islands Adonara, Lomblen and Batoe Tara"; W. P. de Roever, "Geological Investigations in the Southwestern Moetis Region (Netherlands Timor)"; J. H. van Voorthuysen, "Geologische Untersuchungen im Distrikt Amfoan (Nordwest Timor)"; and C. Wanner, "Neue Permische Lamellibranchiaten von Timor."

Volume III, of 380 pages, includes three reports: F. P. van West, "Geological Investigations in the Miomaffo Region (Netherlands Timor)"; D. L. de Bruyne, "Sur la composition et la genèse du bassin central de Timor"; and J. D. de Jong, "Geological Investigations in West Wetar, Lirang and Solor."

Volume IV, 423 pages, is in six sections: J. Heering, "Geological Investigations in East Wetar, Alor and Poera Besar"; C. Wanner, "Neue Beiträge zur Gastropodenfauna des Perm von Timor"; W. P. de Roever, Olivine-Basalts and Their Alkaline Differentiates in the Permian of Timor"; H. A. Brouwer, "Granodioritic Intrusions and Their Metamorphic Aureoles in the Young-Tertiary of Central Flores"; J. D. de Jong, "Hydrothermal Metamorphism in the Lowo Ria Region, Central Flores"; and H. A. Brouwer, "Summary of the Geological Results of the Expedition."

The concluding summary report of 50 pages by Brouwer will be found not only helpful as a general account of the geology of these islands but also particularly valuable for its able treatment of the larger tectonic and magmatic problems and their interrelationships. Few areas of comparable size can contribute more to our understanding of the earth's diastrophic behavior.

R. T. C.

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September 25, 1947

Dr. R. T. Chamberlin, Editor
Journal of Geology
5750 Ellis Ave.
Chicago 37, Illinois

DEAR DR. CHAMBERLIN:

Finland has an excellent and keenly scientific-minded Technical Institute, Teknillinen Korkeakoulu. During the war its library was bombed and totally destroyed.

On my recent trip to Finland for the American Friends Service Committee, I discussed the situation with Dr. Martti Levon, Director of the Institute. He said he would welcome gifts of scientific and technical books and periodicals from America to take the place of those destroyed. In the remarkable efforts for recovery that the Finns are making, the lack of technical library facilities is a very serious handicap. It would be a practical act of friendship to a nation that holds America in high regard if Americans should contribute good technical books and periodicals to this library.

Any such gifts should be marked for the Institute of Technology, Helsinki, and sent to the Legation of Finland, 2144 Wyoming Ave., N.W., Washington, D.C. Dr. K. T. Jutila, the Finnish Minister, will arrange for their being shipped to Finland.

Sincerely yours,

ARTHUR E. MORGAN

AMERICAN FRIENDS SERVICE COMMITTEE
YELLOW SPRINGS, OHIO

INDEX TO VOLUME LV

Accelerated channel erosion in the Cimarron valley in southwestern Kansas (Thad G. McLaughlin), 76

Adams, Sidney. Cited on deformation bands (N. Allen Riley), 469

Africa, South, The Vredefort ring-structure of (Reginald A. Daly), 125

Albritton, C. C., Jr., and J. D. Boon. Cited on the Vredefort ring-structure (Reginald A. Daly), 125, 137, 142

Alden, W. C. Cited on pre-Cambrian monadnocks in Wisconsin (O. E. Gram), 427

Allen, A. T. *The Longview member of the Kingsport formation*, 412

Amebelodon jaw from the Texas panhandle. By J. T. Gregory. Review by Everett C. Olson, 177

Amphibians from the Dockum Triassic of Howard County, Texas. By H. J. Sawin. Review by Everett C. Olson, 177

Anatolian facies, Arabian and, the boundary between (S. W. Tromp), 364
orogenic belt, A tentative classification of the main structural units of the, 362
orogenic belt in southeastern Europe, possible continuation of the, 372
plateau, the central, 372

Anderson, Alfred L. *Drainage diversion in the northern Rocky Mountains of east-central Idaho*, 61
Cited on age of Orofino series, 491
Cited on hornblendite (Clayton H. Johnson), 497, 504
Cited on Idaho batholith, 492
and V. Hammerand. Cited on biotite in contact zones, 504

Anorthite, substances in solid solution (Julian R. Goldsmith), 398

Antarctic seas, The voyage of Captain Bellingshausen in the, 1819–1921. By Frank Debenham (ed.). Review by William Herbert Hobbs, 380

Antarcto-Carboniferous floras (Charles B. Read), 277

Antevs, Ernst. Review of: Dating the past: an introduction to geochronology. By Frederick E. Zeuner, 527
Review of: The Pleistocene period: its climate, chronology and faunal successions. By Frederick E. Zeuner, 446
Review of: Postglacial forest succession, climate, and chronology in the Pacific Northwest. By Henry P. Hansen, 450

Appalachian coal field, types of Pennsylvanian cycles in the (Harold R. Wanless), 239

Appalachian region, microfossil evidence for the division of the Pennsylvanian strata of the (Aureal T. Cross), 306

Appalachian trough, the (F. G. Snyder), 149

Appalachians, southern, Correlation of the erosion surfaces of the (Paul R. Shaffer), 343

Arabian and Anatolian facies, the boundary between (S. W. Tromp), 364

Archipelago, volcanic (A. J. Eardley), 334

Arcto-Carboniferous floras (Charles B. Read), 273

Ashley, G. H. Cited on classification of Clarion beds in Pennsylvania (M. G. Cheney), 209
Cited on the correlation of the Putnam Hill and Vanport limestones, 217

Atwood, W. W. Cited on drainage in western Montana and eastern Idaho (Alfred L. Anderson), 67

Babenroth, D. L., and A. N. Strahler. Cited on faulting (Sheldon Judson), 172

Bailey, E. B. Cited on the Vredefort ring-structure (Reginald A. Daly), 135

Baldwin, E. M. Cited on Lost River Range (Alfred L. Anderson), 68

Bancroft, H. Cited on the Stevens series of Washington (A. J. Eardley), 323

Baraboo quartzite, Structural petrology of the (N. Allen Riley), 453
region, general geologic description of the, 453

Barringer, D. M. Cited on meteorite craters (Reginald A. Daly), 139

Barth, Tom. F. W. *On the geochemical cycle of fluorine*, 420

Barton, Donald. Cited on exfoliation in Egypt (Theodore A. Dodge), 38

Bartrum, J. A. *The rate of rounding of beach boulders*, 514

Basalt series, Omeishan, in southwestern China, On the structure of Psaronius sinensis from the (H. C. Sze), 160

Base of the Cambrian system (Harry E. Wheeler), 153

Basin drainage, in northern Rocky Mountains (Alfred L. Anderson), 73

Bathymetrical map in seven sheets of the Norwegian coastal waters and adjoining areas, The submarine relief off the Norwegian coast, with. By Olaf Holtedahl. Review by Leland Horberg, 58

Bayley, W. S. Cited on rocks in the Pine Creek area (James W. Higgins), 476

Beach boulders, The rate of rounding of (J. A. Bartrum), 514

Beach erosion, Refraction of ocean waves: a process linking underwater topography to (Walter H. Munk and Melvin A. Traylor), 1

Bellemin, George J. Review of: Introduction à la lecture des cartes géologiques. By Antoine Bonte, 57

Berman, Harry. Cited on melilites (Julian R. Goldsmith), 401

Bishopp, D. W. Cited on the Vredefort ring-structure (Reginald A. Daly), 135

Blanco fauna, The. By Grayson E. Meade. Review by Everett C. Olson, 178

Bloomer, Richard R. and Robert O. *The Catoctin formation in central Virginia*, 94

Bloomer, Robert O. *A tectonic intrusion of shale in Rockbridge County, Virginia*, 48

Boncev, E. Cited on overthrusts in Bulgaria (S. W. Tromp), 377

Bondel, Eugene. Cited on the Cimarron River (Thad G. McLaughlin), 81

Bonte, Antoine. Introduction à la lecture des cartes géologiques. Review by George J. Bellemin, 57

Boon, J. D., and C. C. Albritton, Jr. Cited on the Vredefort ring-structure (Reginald A. Daly), 125, 137, 142

Bottom topography and wave refraction (Walter H. Munk and Melvin A. Traylor), 20

Boulders, beach, The rate of rounding of (J. A. Bartrum), 514

Bowen, N. L. Cited on carnegieite and anorthite (Julian R. Goldsmith), 383
Cited on indices of refraction and on inversion points (Robert G. Smalley), 28
Cited on nepheline solid solution with anorthite (Robert G. Smalley), 34
Cited on the system $CaAl_2Si_2O_8$–$NaAlSiO_4$ (Julian R. Goldsmith), 384

Bradfield, H. H. Cited on Pennsylvanian ostracodes (Chalmer L. Cooper), 266

Brock, B. B. Cited on metamorphism of the Shuswap rocks (A. J. Eardley), 332

Brown, Harrison, and Claire Patterson. *The composition of meteoritic matter.* I. *The composition of the silicate phase of stony meteorites*, 405
The composition of meteoritic matter. II. *The composition of iron meteorites and of the metal phase of stony meteorites*, 508

Brown, W. R., and B. N. Cooper. Cited on gravity faults in an anticline (Robert O. Bloomer), 50

Bryan, Kirk. Cited on accelerated channel erosion (Thad G. McLaughlin), 89

Bucher, W. H. (Cited on use of the stereographic net in determination of joint direction (James W. Higgins), 484

Buddington, A. F. Cited on structural trends of the Triassic conglomerate (A. J. Eardley), 334

Buerger, M. J. Cited on soda-bearing tridymites (Julian R. Goldsmith), 384

Butts, Charles. Cited on pre-Cambrian deposits (F. G. Snyder), 152

Cailleux, André. Cited on the chalk in the lower Seine region (Sheldon Judson), 171

Cairnes, C. E. Cited on the Nisconlith series of British Columbia (A. J. Eardley), 323
Cited on metamorphism of the Shuswap rocks in British Columbia, 331

California, La Jolla. *Refraction of ocean waves: a process linking underwater topography to beach erosion* (Walter H. Munk and Melvin A. Traylor), 1

Cambrian system, Base of the (Harry E. Wheeler), 153
proposed base of the (Harry E. Wheeler), 157

Campbell, M. R., and C. W. Hayes. Cited on peneplains in the southern Appalachians (Paul R. Shaffer), 343

Carnegieite (Robert G. Smalley), 27
substances in solid solution (Julian R. Goldsmith), 400

Catoctin formation in central Virginia, The (Robert O. Bloomer and Richard R. Bloomer), 94
age of, 102
classification of, 99
distribution and stratigraphic relations of, 94
origin of, 100
structure and petrography of, 97
thickness of, 101

Cavern, A fault-plane (E. L. Krinitzsky), 107

Chamberlin, Rollin T. Review of: Geological expedition to the Lesser Sunda Islands under the leadership of H. A. Brouwer, 531
Review of: Geology and ground-water resources of the island of Hawaii. By H. T. Stearns and G. A. Macdonald, 378
Review of: German-English science dictionary. By Louis De Vries, 60
Review of: India, Part I: Physical basis of geography of India. By H. L. Chhibber, 60
Review of: Journal of glaciology, 180
Review of: On major tectonic forms of China. By T. K. Huang, 59

Channel erosion in the Cimarron Valley in southwestern Kansas, Accelerated (Thad G. McLaughlin), 76

Chapman, Carleton A., and George K. Schweitzer. *Electrode potentials and free-energy changes in geology*, 43

Chemical weathering, An example of exfoliation caused by (Theodore A. Dodge), 38

Cheney, M. G. *Pennsylvanian classification and correlation problems in north-central Texas*, 202

Chhibber, H. L. India, Part I: Physical basis of geography of India. Review by Rollin T. Chamberlin, 60

China, On major tectonic forms. of. By T. K. Huang. Review by Rollin T. Chamberlin, 59
southwestern, On the structure of Psaronius sinensis from the Omeishan basalt series in (H. C. Sze), 160

Cimarron Valley in southwestern Kansas, Accelerated channel erosion in the (Thad G. McLaughlin), 76

Clarke, F. W. Cited on amount of fluorine in the earth's crust (Tom. F. W. Barth), 420

Coal beds, Plant microfossils in correlation of (R. M. Kosanke), 280

Coastline of England and Wales, The. By J. A. Steers. Review by Leland Horberg, 378

Cole, W. S. Cited on surfaces in the northern Appalachians (Paul R. Shaffer), 343

Colorado, western, Upright trunks of Neocalamites from the Upper Triassic of (Edward L. Holt), 511

Composition of meteoritic matter, The. I. *The composition of the silicate phase of stony meteorites* (Harrison Brown and Claire Patterson), 405
II. *The composition of iron meteorites and of the metal phase of stony meteorites* (Harrison Brown and Claire Patterson), 508

Condra, G. E., and C. O. Dunbar. Cited on Pennsylvanian brachiopods (J. Marvin Weller), 258

Conodonts (Chalmer L. Cooper), 269

Contributions to geology, 1944. Review by J. Marvin Weller and Everett C. Olson, 176

Cooke, C. Wythe. Geology of Florida. Review by Garald G. Parker, 120

Cooper, B. N., and W. R. Brown. Cited on gravity faults in an anticline (Robert O. Bloomer), 50

Cooper, Chalmer L. *Role of microfossils in interregional Pennsylvanian correlations*, 261
Pennsylvanian ostracodes of Illinois. Review by Betty Kellett Nadeau, 176
Cited on Pennsylvanian ostracodes (J. Marvin Weller), 258

Cordilleran geosyncline and related orogeny, Paleozoic (A. J. Eardley), 309

Cordilleran trough (F. G. Snyder), 147

Correlation of the erosion surfaces of the southern Appalachians (Paul R. Shaffer), 343

Cotton, C. A. Geomorphology: an introduction to the study of landforms. Review by Leland Horberg, 378

Craig, L. C.; P. B. King; H. W. Ferguson; and John Rodgers. Cited on the base of the Cambrian (F. G. Snyder), 149

Cross, Aureal T. *Spore floras of the Pennsylvanian of West Virginia and Kentucky*, 285
Cited on floristic changes (M. G. Cheney), 217

Crystalline phases, properties of (Robert G. Smalley), 27
β-Al_2O_3 (Julian R. Goldsmith), 384
$Ca_2Al_2SiO_7$, 383
$CaAl_2Si_2O_8$, 382
$NaAlSiO_4$, 383

Crystallization, courses of (Julian R. Goldsmith), 395

Cummins, W. F. Cited on Brazos River Valley section (M. G. Cheney), 218

Current direction, Primary lineation in fluvial sandstones, a criterion of (W. Lee Stokes), 52

Daly, Reginald A. *The Vredefort ring-structure of South Africa*, 125

Dappert, J. W. Cited on the Cimarron River (Thad G. McLaughlin), 84

Dapples, E. C., and J. F. Rominger. Cited on current direction (W. Lee Stokes), 52

Darwin, Charles, and the voyage of the Beagle. Edited by Nora Barlow. Review by J. Marvin Weller, 452

Darwin, Discovery of letters by Lyell and (Henry P. Zuidema), 439

Dating the past: an introduction to geochronology. By Frederick E. Zeuner. Review by Ernst Antevs, 527

Daugherty, L. H. Cited on *Neocalamites* in Arizona (Edward L. Holt), 513

Davis, W. M. Cited on erosion surfaces of the northern Appalachians (Paul R. Shaffer), 343
Cited on origin of caves (E. L. Krinitzsky), 114

Dean, E. M. Cited on the Cimarron River (Thad G. McLaughlin), 82

Debenham, Frank (ed.). The voyage of Captain Bellingshausen in the Antarctic seas 1819–1821. Review by William Herbert Hobbs, 380

Decapod crustaceans from the Cretaceous of Texas. By H. B. Stenzel. Review by J. Marvin Weller, 177

Decker, Charles E. The Wilberns Upper Cambrian graptolites from Mason, Texas.
Review by J. Marvin Weller, 176

Deformation bands (N. Allen Riley), 468
lamellae, 461

Deformational structures, major, of the Pine Creek area (James W. Higgins), 479

Deiss, C. F. Cited on Cascadis (A. J. Eardley), 334
Cited on classic Cambrian sections of British Columbia and Alberta (F. G. Snyder), 147, 148

Depth, relative, concept of (Walter H. Munk and Melvin A. Traylor), 21

Description of Lower Pennsylvanian corals from Texas and adjacent states. By Raymond C. Moore and Russell M. Jeffords. Review by J. Marvin Weller, 176

Devitrification, of pre-Cambrian rocks in Wisconsin (O. E. Gram), 430

De Vries, Louis. German-English science dictionary. Review by Rollin T. Chamberlin, 60

de Wyckerslooth, P. Cited on chromium deposits in the Varder zone (S. W. Tromp), 374

Diabase dikes of the Franklin Furnace, New Jersey, quadrangle (Charles Milton), 522
field relations of the, 523
location of, 522

Diabases, of the Franklin Furnace, petrography of the (Charles Milton), 524

Dickinson County, Michigan, Structural petrology of the Pine Creek area (James W. Higgins), 476

Dikes, Diabase, of the Franklin Furnace, New Jersey, quadrangle (Charles Milton), 522

Dilation, hypothesis, the Hollingsworth adit and the (Theodore A. Dodge), 40
as a prerequisite of exfoliation, 38

Distribution and correlation of Pennsylvanian rocks in late Paleozoic sedimentary basins of northern New Mexico (Charles B. Read and Gordon H. Wood), 220

Diversion, Drainage, in the northern Rocky Mountains of east-central Idaho (Alfred L. Anderson), 61

Dodge, Theodore A. *An example of exfoliation caused by chemical weathering*, 38

Dolmage, Victor. Cited on metamorphosed Carboniferous rocks (A. J. Eardley), 327

Dosch, E. F., C. W. Thornthwaite, and C. F. S. Scharpe. Cited on accelerated channel erosion (Thad G. McLaughlin), 89

Drainage, development, history of (Alfred L. Anderson), 73
diversion, dating the, 73
diversion, evidence of, 71
diversion in the northern Rocky Mountains of east-central Idaho, 61
early northeast, evidence of, in northern Rocky Mountains, 68
pattern in northeastern Minnesota, The influence of geologic structure on the (Karl ver Steeg), 353

Duluth gabbro (Keweenawan) (Karl ver Steeg), 360

Dunbar, C. O., and G. E. Condra. Cited on Pennsylvanian brachiopods (J. Marvin Weller), 258
L. G. Henbest and J. Marvin Weller. Cited on fusulinid correlation (M. G. Cheney), 217
and Schuchert, Charles. Cited on *Calamites* and *Sigillaria* at Joggins, Nova Scotia (Edward L. Holt), 511
and Schuchert, Charles. Cited on the Lipalian interval (F. G. Snyder), 150

Dutton, C. E., and Lamey, C. A. Cited on faults in the Menominee iron-bearing district (James W. Higgins), 480

Eardley, A. J. *Paleozoic Cordilleran geosyncline and related orogeny*, 309

Ecology, marine, as related to paleontology, Report of the committee on. By Harry S. Ladd, chairman. Review by J. Marvin Weller, 180

Electrode potentials and free-energy changes in geology (Carleton A. Chapman and George K. Schweitzer), 43
geological applications of, 45
significance of, 44

Ellison, Samuel. Cited on Pennsylvanian conodonts (Chalmer L. Cooper), 269

Ely greenstone (Keewatin) (Karl ver Steeg), 356

Emmons, Ebenezer. Cited on origin of the Raleigh graphite (John W. Harrington), 519

Equilibrium relationships, high-temperature. *The system* $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Robert G. Smalley), 27

Erosion, accelerated, causes of (Thad G. McLaughlin), 89
Accelerated channel, in the Cimarron Valley in southwestern Kansas, 76
beach, Refraction of ocean waves: a process linking underwater topography to (Walter H. Munk and Melvin A. Traylor), 1
stratigraphy and the, 351
surfaces of the southern Appalachians, Correlation of (Paul R. Shaffer), 343
in tributary streams, 86
wave refraction and, 23

Europe, southeastern, possible continuation of the Anatolian orogenic belt in (S. W. Tromp), 373

Evans, Glen L., and Grayson E. Meade. Quaternary of the Texas high plains. Review by Everett C. Olson, 177

Example of exfoliation caused by chemical weathering, An (Theodore A. Dodge), 38

Exfoliation, dilation as a prerequisite of (Theodore A. Dodge), 38
An example of, caused by chemical weathering, 38
postulated causes of, 38

Fairbairn, H. W. Cited on deformation lamellae (N. Allen Riley), 461
Cited on development of *a*-maxima in slickensides (James W. Higgins), 488

Fanning, Leonard M. Editor of: Our oil resources. Review by A. I. Levorsen, 55

Farmin, Rollin. Cited on cause of exfoliation (Theodore A. Dodge), 38

Fault-plane cavern, A (E. L. Krinitzsky), 107

Faunal studies (J. Marvin Weller), 255

Fellows, R. E. Cited on "crushed qaurtz" texture (N. Allen Riley), 460

Ferguson, H. W., P. B. King, L. C. Craig, and John Rodgers. Cited on the base of the Cambrian (F. G. Snyder), 149

Fisk, Harold N. Geological investigation of the alluvial valley of the lower Mississippi River. Review by William Herbert Hobbs, 378

Flint, Richard Foster. Glacial geology and the Pleistocene epoch. Review by Leland Horberg, 530

Floral provinces, Pennsylvanian floral zones and (Charles B. Read), 271

Floral zones, Pennsylvanian, and floral provinces, 271

Floras, of Pennsylvanian age (Charles B. Read), 273
Spore, of the Pennsylvanian of West Virginia and Kentucky (Aureal T. Cross), 285

Florida, Geology of. By C. Wythe Cooke. Review by Garald G. Parker, 120

Fluorine concentration in the lithosphere, the average (Tom. F. W. Barth), 420
On the geochemical cycle of, 420
in the Oslo region, 422
in rocks and waters from Iceland, 423
in sea water, 421

Fluvial sandstones, Primary lineation in, a criterion of current direction (W. Lee Stokes), 52

Foraminifera (Chalmer L. Cooper), 262
Smaller, in the Marble Falls, Smithwick, and lower Strawn strata around the Llano uplift in Texas. By Helen Jeanne Plummer. Review by J. Marvin Weller, 176

Franklin Furnace, New Jersey, quadrangle, Diabase dikes of the (Charles Milton), 522

Free-energy changes (Carleton A. Chapman and George K. Schweitzer), 45
in geology, Electrode potentials and, 43

Frye, J. C. Cited on Crooked Creek (Thad G. McLaughlin), 88
and C. W. Hibbard. Cited on high-level terraces (Thad G. McLaughlin), 78

Furcron, A. S. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94, 99, 101, 102

Gardner, J. H. Cited on development of caves (E. L. Krinitzsky), 116, 118

Geanticlines, geosynclines, and forelands, position of, in New Mexico (Charles B. Read and Gordon H. Wood), 225
Gehlenite (Robert G. Smalley), 29
substances in solid solution (Julian R. Goldsmith), 400
Geiger, H. R., and Arthur Keith. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94
Genus, specific characteristics and possible variations in a (Aureal T. Cross), 293
Geochemical cycle of fluorine, On the (Tom. F. W. Barth), 420
Geologic structure, The influence of, on the drainage pattern in northeastern Minnesota (Karl ver Steeg), 353
Geological applications of electrode potentials (Carleton A. Chapman and George K. Schweitzer), 45
Geological expedition to the Lesser Sunda Islands under the leadership of H. A. Brouwer. Review by Rollin T. Chamberlin, 531
Geological investigation of the alluvial valley of the lower Mississippi River. By Harold N. Fisk. Review by William Herbert Hobbs, 378
Geology, of Florida. By C. Wythe Cooke. Review by Garald G. Parker, 120
and geologic history, of southwestern Kansas (Thad G. McLaughlin), 78
and ground water resources of the island of Hawaii. By H. T. Stearns and G. A. Macdonald. Review by Rollin T. Chamberlin, 378
Geomorphology: an introduction to the study of landforms. By C. A. Cotton. Review by Leland Horberg, 378
Geosyncline, Paleozoic Cordilleran, and related orogeny (A. J. Eardley), 309
German-English science dictionary. By Louis De Vries. Review by Rollin T. Chamberlin, 60
Gilluly, James. Cited on the Burnt River schist (A. J. Eardley), 320
Cited on metamorphism of the Shuswap rocks (A. J. Eardley), 332
and J. B. Reeside. Cited on Triassic red beds (Edward L. Holt), 512
Girty, G. H. Cited on Wewoka fauna of Oklahoma (J. Marvin Weller), 257
Glacial geology and the Pleistocene epoch. By Richard Foster Flint. Review by Leland Horberg, 530
Glaciation, effects of (Karl ver Steeg), 360
Glaciology, Journal of. Review by Rollin T. Chamberlin, 180
Goldschmidt, V. M. Cited on cosmic abundance of nuclear species (Harrison Brown and Claire Patterson), 405
Goldsmith, Julian R. *The system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$*, 381
Goodwins Ferry cave, deposits in (E. L. Krinitzsky), 114
description of, 109
A fault-plane cavern, 107
hydrology of, 112
origin of, 114
stratigraphy of, 107
structure of, 109
Gould, C. N. Cited on the Cimarron River (Thad G. McLaughlin), 82
Grabau, A. W. Cited on the Sinian system (Harry E. Wheeler), 159
Grain elongation, dimensional (N. Allen Riley), 473
Gram, O. E. *Tectonic features of the Utley metarhyolite*, 427
Graphite, Raleigh, The origin and importance of the (John W. Harrington), 516
Great Basin area, time-stratigraphic relationships in the southern (Harry E. Wheeler), 155
Great Basin region. *Base of the Cambrian system* (Harry E. Wheeler), 153
Gregg, Josiah. Cited on the Cimarron River (Thad G. McLaughlin), 81
Gregory, J. T. An *Amebelodon* jaw from the Texas panhandle. Review by Everett C. Olson, 177
Osteology and relationships of *Trilophosaurus*. Review by Everett C. Olson, 177
Grout, F. F. Cited on the Duluth gabbro (Karl ver Steeg), 360
Cited on glassy textures (O. E. Gram), 430
Guide fossils (J. Marvin Weller), 257
Gummer, W. K. Cited on extent of field of β-Al_2O_3 (Julian R. Goldsmith), 385
Gunflint iron formation (Animikie) (Karl ver Steeg), 356

Hall, A. L., and G. A. F. Molengraaff. Cited on the stratigraphy of the Vredefort ring-structure (Reginald A. Daly), 126, 128, 132, 133
Hammerand, V., and A. L. Anderson. Cited on biotite in contact zones (Clayton H. Johnson), 504
Hansen, Henry P. Postglacial forest succession, climate, and chronology in the Pacific Northwest. Review by Ernst Antevs, 450
Harrington, John W. *The origin and importance of the Raleigh graphite*, 516
Haworth, Erasmus. Cited on the Cimarron River (Thad G. McLaughlin), 81
Hayes, C. W., and M. R. Campbell. Cited on peneplains of the southern Appalachians (Paul R. Shaffer), 343
Hazzard, J. C. Cited on the *Olenellus* zone (Harry E. Wheeler), 155, 157
Cited on stratigraphic section at Nopah–Resting Springs, California (F. G. Snyder), 147
Headwater drainage, in northern Rocky Mountains (Alfred L. Anderson), 68
Hedvall, J. A., and Sjöman. Cited on diffusion of Fe_2O_3 into the quartz lattice (N. Allen Riley), 467
Henbest, Lloyd G., C. O. Dunbar, and J. Marvin Weller. Cited on fusulinid correlation (M .G. Cheney), 217
Hibbard, C. W., and J. C. Frye. Cited on high-level terrace (Thad G. McLaughlin), 78

Hietanen, Anna. Cited on deformation lamellae (N. Allen Riley), 466
Cited on undulatory extinction, 459, 468
Higgins, James W. *Structural petrology of the Pine Creek area, Dickinson County, Michigan*, 476
Hinds, N. E. A. Cited on Beltian sediments (F. G. Snyder), 148
Cited on the section in the southern Klamath Mountains (A. J. Eardley), 316
Hirmer, M. Cited on *Psaronius quadrangulus* (H. C. Sze), 163
Hobbs, William Herbert. Review of: Geological investigation of the alluvial valley of the lower Mississippi River. By Harold N. Fisk, 378
Review of: The voyage of Captain Bellingshausen in the Antarctic seas 1819–1821. Edited by Grank Debenham, 380
Cited on andesites and basalts of the Klamath and Alexander troughs (A. J. Eardley), 335
Hollingsworth adit and the dilation hypothesis, the (Theodore A. Dodge), 40
Hollingworth, S. E., and J. H. Taylor. Cited on cambers (Sheldon Judson), 170
Holmes, A. Cited on pseudo-tachylyte (Reginald A Daly), 132
Holt, Edward L. *Upright trunks of Neocalamites from the Upper Triassic of western Colorado*, 511
Holtedahl, Olaf. The submarine relief off the Norwegian coast with bathymetrical map in seven sheets of the Norwegian coastal waters and adjoining areas. Review by Leland Horberg, 58
Horberg, Leland. Review of: The coastline of England and Wales. By. J. A. Steers, 378
Review of: Geomorphology: an introduction to the study of landforms. By C. A. Cotton, 378
Review of: Glacial geology and the Pleistocene epoch. By Richard Foster Flint, 530
Review of: The submarine relief off the Norwegian coast with bathymetrical map in seven sheets of the Norwegian coastal waters and adjoining areas. By Olaf Holtedahl, 58
Huang, T. K. On major tectonic forms of China. Review by Rollin T. Chamberlin, 59
Cited on age of *Gigantopteris* flora (H. C. Sze), 164
Hubbert, M. K. Cited on overtilting of caliche (Reginald A. Daly), 140, 145
Hunt, C. B. Cited on faulting (Sheldon Judson), 172
Hutchison, William E. Cited on the Cimarron River (Thad G. McLaughlin), 84

Iceland, fluorine in rocks and waters from (Tom. F. W. Barth), 423
Idaho, east-central, Drainage diversion in the northern Rocky Mountains of (Alfred L. Anderson), 61
Orofino region, Igneous metamorphism in the (Clayton H. Johnson), 490
Igneous metamorphism in the Orofino region, Idaho (Clayton H. Johnson), 490
Igneous rocks, in the Utley area (O. E. Gram), 427
Illinois. *Plant microfossils in correlation of coal beds* (R. M. Kosanke), 280
India, Part I: Physical basis of geography of India. By H. L. Chhibber. Review by Rollin T. Chamberlin, 60
Influence of geologic structure on the drainage pattern in northeastern Minnesota, The (Karl ver Steeg), 353
Ingerson, Earl. Cited on ripple marks (W. Lee Stokes), 52
and O. F. Tuttle. Cited on deformation lamellae (N. Allen Riley), 461, 462, 466
Interregional correlation, principles of (Aureal T. Cross), 293
Introduction à la lecture des cartes géologiques. By Antoine Bonte. Review by George J. Bellemin, 57
Introduction to symposium on Pennsylvanian problems (Harold R. Wanless), 184
Intrusion, A tectonic, of shale in Rockbridge County, Virginia (Robert O. Bloomer), 48
Intrusives, the Rove slate and associated, topography of the area underlain by (Karl ver Steeg), 358
Invertebrates in Pennsylvanian correlations (J. Marvin Weller), 254
Izmit-Erzurum line, the (S. W. Tromp), 365

Jeffords, Russell M., and R. C. Moore. Description of Lower Pennsylvanian corals from Texas and adjacent states. Review by J. Marvin Weller, 176
Johnson, Clayton H. *Igneous metamorphism in the Orofino region, Idaho*, 490
Johnson, W. D. Cited on the Cimarron River (Thad G. McLaughlin), 81
Jointing, of the rocks in the Pine Creek area (James W. Higgins), 482
Joly, J. Cited on primeval sea (Tom. F. W. Barth), 421
Jonas, A. I. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94, 102
Judson, Sheldon. *Large-scale superficial structures—a discussion*, 168

Kansas, southwestern, Accelerated channel erosion in the Cimarron valley in (Thad G. McLaughlin), 76
Keith, Arthur. Cited on base of the Cambrian (F. G. Snyder), 149
Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94, 99, 100, 101, 102
and H. R. Geiger. Cited on the Catoctin formation, 94
Kellett, Betty. Cited on Pennsylvanian and Permian ostracodes (Chalmer L. Cooper), 266
Kentucky, West Virginia and, Spore floras of the Pennsylvanian of (Aureal T. Cross), 285
Kessler, Paul. Cited on the Triassic Voltzian sandstone (Sheldon Judson), 171
King, P. B., H. W. Ferguson, L. C. Craig, and John Rodgers. Cited on the base of the Cambrian (F. G. Snyder), 149

Kingsport formation, The Longview member of the (Arthur T. Allen), 412
Kirkham, V. R. D. Cited on origin of valley depressions (Alfred L. Anderson), 68
Knopf, A. Cited on the Calaveras in the Mother Lode area (A. J. Eardley), 330
and L. G. Westgate. Cited on thickness of sediments, 314
Knox dolomite. *The Longview member of the Kingsport formation* (Arthur T. Allen), 412
Koenigsberger, J. G., and E. Hungerer. Cited on secondary folds in bending of prisms (N. Allen Riley), 475
Kosanke, R. M. *Plant microfossils in correlation of coal beds*, 280
Krauskopf, D. B. Cited on the Anarchist series (A. J. Eardley), 323
Krinitzsky, E. L. *A fault-plane cavern*, 107
Krumbein, W. C. Cited on the effect of wave refraction of sediment transport (Walter H. Munk and Melvin A. Traylor), 22
Krynine, P. D. Cited on origin of sedimentary orthoquartzites and quartzose sandstones (A. J. Eardley), 340
Cited on the tectonic significance of arkoses, 339

Ladd, Harry S. Report of the committee on marine ecology as related to paleontology. Review by J. Marvin Weller, 180
Lahee, F. H. Cited on criteria for determining direction of movement of currents (W. Lee Stokes), 52
La Jolla, California, wave refraction along the beach north of (Walter H. Munk and Melvin A. Traylor), 8
Lamey, C. A. Cited on rocks in the Pine Creek area (James W. Higgins), 476
and C. E. Dutton. Cited on faults in the Menominee iron-bearing district, 480
Large-scale superficial structures—a discussion (Sheldon Judson), 168
Leith, Andrew. Cited on the division of the Baraboo Huronian sediments (N. Allen Riley), 455, 456
Levorsen, A. I. Review of: Our oil resources. Edited by Leonard M. Fanning, 55
Lineation, Primary, in fluvial sandstones, a criterion of current direction (W. Lee Stokes), 52
Lipalian interval, The problem of the (F. G. Snyder), 146
Liquidus relations, the, for the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Robert G. Smalley), 32
Lithology, Pennsylvanian, Regional variations in (Harold R. Wanless), 237
Lithosphere, the average fluorine concentration in the (Tom. F. W. Barth), 420
Llano region, Summary of classification of the Pennsylvanian formations of Texas, with special reference to the Lower Pennsylvanian of the (Frederick B. Plummer), 193
Longview, dolomite, stratigraphy of the (Arthur T. Allen), 414
member of the Kingsport formation, The, 412
Lowe, Kurt E. Cited on lineation of axes of grain elongation (N. Allen Riley), 473
Lowman, Shepard W. Cited on structural history of the Mid-Continent region (M. G. Cheney), 213
Lyell and Darwin, Discovery of letters by (Henry P. Zuidema), 439

Macdonald, G. A., and H. F. Stearns. Geology and ground-water resources of the island of Hawaii. Review by Rollin T. Chamberlin, 378
McKee, E. D. Cited on the *Olenellus* horizon (Harry E. Wheeler), 155
McLaughlin, Thad G. *Accelerated channel erosion in the Cimarron Valley in southwestern Kansas*, 76
Malott, C. A. Cited on development of cave systems (E. L. Krinitzsky), 116
Mansfield, G. R. Cited on Upper Mississippian beds in Idaho (A. J. Eardley), 314
Maree, B. D. Cited on the Vredefort ring-structure (Reginald A. Daly), 125, 128, 134, 135
Mead, W. J. Cited on horizontal shearing movements (O. E. Gram), 438
Meade, Grayson E. The Blanco fauna. Review by Everett C. Olson, 178
and Glen L. Evans. Quaternary of the Texas high plains. Review by Everett C. Olson, 177
Meinzer, O. E. Cited on means of valley depression (Alfred L. Anderson), 67
Merriam, C. W. Cited on Devonian beds in Nevada (A. J. Eardley), 313
Merrill, G. P. Cited on average composition of stone meteorites (Harrison Brown and Claire Patterson), 405
Merwin, H. E., and G. A. Rankin. Cited on β-Al_2O_3 (Julian R. Goldsmith), 384
Metamorphism, Igneous, in the Orofino region, Idaho (Clayton H. Johnson), 490
Meta-rhyolite, Utley, Tectonic features of the (O. E. Gram), 427
Meteorite scar, does the Vredefort ring-structure represent a? (Reginald A. Daly), 137
Meteoritic matter, The composition of. I. *The composition of the silicate phase of stony meteorites* (Harrison Brown and Claire Patterson), 405
II. *The composition of iron meteorites and of the metal phase of stony meteorites* (Harrison Brown and Claire Patterson), 508
Meteorites, stony, The Composition of meteoritic matter. I. *The composition of the silicate phase of* (Harrison Brown and Claire Patterson), 405
Meteorology and wave refraction (Walter H. Munk and Melvin A. Traylor), 19
Michigan, and eastern interior basins, types of Pennsylvanian cycles in the (Harold R. Wanless), 241
Pine Creek area, Dickinson County, Structural petrology of the (James W. Higgins), 476

Microfossil evidence of the division of the Pennsylvanian strata of the Appalachian region (Aureal T. Cross), 306

Microfossils, identification and classification of (Aureal T. Cross), 290
Plant, in correlation of coal beds (R. M. Kosanke), 280
Role of, in interregional Pennsylvanian correlations (Chalmer L. Cooper), 261

Microfractures (N. Allen Riley), 470

Microspore assemblages, discussion of the (Aureal T. Cross), 305

Microstructure, details of the, of the Baraboo quartzite (N. Allen Riley), 461

Mid-Continent region, types of Pennsylvanian cycles in the (Harold R. Wanless), 243

Milton, Charles, *Diabase dikes of the Franklin Furnace, New Jersey, quadrangle*, 522

Minnesota, northeastern, The influence of geologic structure on the drainage pattern in (Karl ver Steeg), 353

Molengraaff, G. A. F., and A. L. Hall. Cited on the stratigraphy of the Vredefort ring-structure (Reginald A. Daly), 126, 128, 132, 133

Moore, R. C. Cited on Pennsylvanian cycles (Harold R. Wanless), 237, 238, 244
Cited on trough sediments (Harry E. Wheeler), 158
and R. M. Jeffords. Description of Lower Pennsylvanian corals from Texas and adjacent states. Review by J. Marvin Weller, 176

Muller, S. W., and H. G. Schenck. Cited on lithologic boundaries (Harry E. Wheeler), 153

Munk, Walter H., and Melvin A. Traylor. *Refraction of ocean waves: a process linking underwater topography to beach erosion*, 1

Nadeau, Betty Kellett. Review of: Pennsylvanian ostracodes of Illinois. By Chalmer L. Cooper, 176

Nel, L. T. Cited on the stratigraphy of the Vredefort ring-structure (Reginald A. Daly), 126, 130, 131, 134, 144

Nelson, W. A. Cited on the Crab Orchard anticline (Paul R. Shaffer), 351

Neocalamites, Upright trunks of, from the Upper Triassic of western Colorado (Edward L. Holt), 511
conditions of sedimentation, 513
stratigraphy and mode of occurrence of the plants, 512

Nepheline (Robert G. Smalley), 28
Nepheline-carnegieite inversion, the (Robert G. Smalley), 33
substances in solid solution (Julian R. Goldsmith), 399

New Jersey, Franklin Furnace, quadrangle, Diabase dikes of the (Charles Milton), 522

New Mexico, northern, Distribution and correlation of Pennsylvanian rocks in late Paleozoic sedimentary basins of (Charles B. Read and Gordon H. Wood), 220

Newman, Mark H. Cited on mines in the Mascot district (Arthur T. Allen), 416

Noddack, I. and W. Cited on minor constituents of stony meteorites (Harrison Brown and Claire Patterson), 411

Nolan, T. B. Cited on Pennsylvanian sediments (A. J. Eardley), 316
Cited on the Wood Canyon formation (Harry E. Wheeler), 155

North Carolina. *The origin and importance of the Raleigh graphite* (John W. Harrington), 516

Norwegian coast, The submarine relief off the, with bathymetrical map in seven sheets of the Norwegian coastal waters and adjoining areas. By Olaf Holtedahl. Review by Leland Horberg, 58

Ocean waves, Refraction of: a process linking underwater topography to beach erosion (Walter H. Munk and Melvin A. Traylor), 1

Öpik, E. Cited on behavior of rock matter (Reginald A. Daly), 145

Oil resources, Our. Edited by Leonard M. Fanning. Review by A. I. Levorsen, 55

Olson, Everett C. Review of: An *Amebelodon* jaw from the Texas panhandle. By J. T. Gregory, 177
Review of: Amphibians from the Dockum Triassic of Howard County, Texas. By H. J. Sawin, 177
Review of: The Blanco fauna. By Grayson E. Meade, 178
Review of: Osteology and relationships of *Trilophosaurus*. By Joseph T. Gregory, 177
Review of: Quarternary of the Texas high plains. By Glen L. Evans and Grayson E. Meade, 177
and J. Marvin Weller. Review of: Contributions to geology, 1944, 176

Omeishan basalt series in southwestern China, On the structure of Psaronius sinensis from the (H. C. Sze), 160

Origin and importance of the Raleigh graphite, The (John W. Harrington), 516

Orofino region, Idaho, Igneous metamorphism in the (Clayton H. Johnson), 490
biotite–quartz diorite gneiss, 498
general geologic setting, 490
granodiorite gneiss, 499
hornblende–quartz diorite gneiss, 494
hornblendite, 497
Idaho batholith–quartz diorite, 492
igneous rocks, 492
marble, 502
metasedimentary rocks, 494
petrogenesis, 503
petrography, 491
ultra-basic rock—dunite, 492

Orogenic belt, Anatolian, A tentative classification of the main structural units of the (S. W. Tromp), 362

Orogeny, Paleozoic Cordilleran geosyncline and related (A. J. Eardley), 309
Paleozoic, in the Pacific trough, evidence of (A. J. Eardley), 328

Oslo region, fluorine in the (Tom. F. W. Barth), 422

Osteology and relationships of *Trilophosaurus*. By Joseph T. Gregory. Review by Everett C. Olson, 177
Ostracodes (Chalmer L. Cooper), 265

Pacific trough (A. J. Eardley), 316
evidence of Paleozoic orogeny in the, 328
Paleozoic, late, sedimentary basins of northern New Mexico, Distribution and correlation of Pennsylvanian rocks in (Charles B. Read and Gordon H. Wood), 220
Paleozoic Cordilleran geosyncline and related orogeny (A. J. Eardley), 309
Paleozoic orogeny in the Pacific trough, evidence of (A. J. Eardley), 328
Parker, Garald G. Review of: Geology of Florida. By C. Wythe Cooke, 120
Parker, H. N. Cited on the Cimarron River (Thad G. McLaughlin), 82
Patterson, Claire, and Harrison Brown. *The composition of meteoritic matter*. I. *The composition of the silicate phase of stony meteorites*, 405
The composition of meteoritic matter. II. *The composition of iron meteorites and of the metal phase of stony meteorites*, 508
Pennsylvanian, beds, persistence of (Harold R. Wanless), 248
classification and correlation problems in north-central Texas (M. G. Cheney), 202
correlations, interregional, Role of microfossils in (Chalmer L. Cooper), 261
correlations, Invertebrates in (J. Marvin Weller), 254
cycles, types of, in the Appalachian coal field (Harold R. Wanless), 239
cycles, types of, in the Michigan and eastern interior basins, 241
cycles, types of, in the Mid-Continent region, 243
cycles, types of, in the Rocky Mountains, 245
floral zones and floral provinces (Charles B. Read), 271
floras, standard sequence of, 273
formations of Texas, Summary of classification of, with special reference to the Lower Pennsylvanian of the Llano region (Frederick B. Plummer), 193
geology of a part of the southern Appalachian coal field. By Harold R. Wanless. Review by J. Marvin Weller, 181
lithology, graphical representation of variations in (Harold R. Wanless), 248
lithology, Regional variations in (Harold R. Wanless), 237
Lower, of the Llano region, Summary of classification of the Pennsylvanian formations of Texas, with special reference to (Frederick B. Plummer), 193
ostracodes of Illinois. By Chalmer L. Cooper. Review by Betty Kellett Nadeau, 178
problems, Introduction to symposium on (Harold R. Wanless), 184
rocks in late Paleozoic sedimentary basins of northern New Mexico, Distribution and correlation of (Charles B. Read and Gordon H. Wood), 220
series, nature of sedimentation in the, 223
series of northern New Mexico, major subdivisions of the, 223
series of northern New Mexico, minor subdivisions of the, 223
strata of the Appalachian region, microfossil evidence for the division of the (Aureal T. Cross), 306
studies, techniques applied to (Harold R, Wanless), 190
of West Virginia and Kentucky, Spore floras of the (Aureal T. Cross), 285
Petrofabrics, of Pine Creek area (James W. Higgins), 485
Petrogenesis, of igneous rocks (Clayton H. Johnson), 505
of metasedimentary rocks, 503
Petrography of the diabases, of Franklin Furnace (Charles Milton), 524
of rocks in the Orofino region, Idaho (Clayton H. Johnson), 491
Petrology, experimental. *The system* $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Robert G. Smalley), 27
Structural, of the Baraboo quartzite (N. Allen Riley), 453
Structural, of the Pine Creek area, Dickinson County, Michigan (James W. Higgins), 476
Pettijohn, F. J. Cited on rocks of the Browning Creek area (James W. Higgins), 481
Cited on the rocks of the Pine Creek area, 476, 477
Phillips, F. C. Cited on composite diagrams in structural petrology (James W. Higgins), 486
Pine Creek area, Dickinson County, Michigan, Structural petrology of the (James W. Higgins), 476
general geology, 476
jointing, 482
major deformational structures, 479
petrofabrics, 485
primary structural unconformity, 477
Plane $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ plus 10 per cent $CaSiO_3$, the (Julian R. Goldsmith), 394
Plant microfossils in correlation of coal beds (R. M. Kosanke), 280
Pleistocene period, The: its climate, chronology and faunal successions. By Frederick E. Zeuner. Review by Ernst Antevs, 446
Plummer, F. B. Stratigraphy of the Lower Pennsylvanian coral-bearing strata of Texas. Review by J. Marvin Weller, 176
Plummer, Frederick B. *Summary of classification of the Pennsylvanian formations of Texas, with special reference to the Lower Pennsylvanian of the Llano region*, 193
Cited on Morrowan rocks in Colorado (M. G. Cheney), 211
Plummer, Helen Jeanne. Smaller formainifera in the Marble Falls, Smithwick, and Lower Strawn beds around the Llano uplift in Texas. Review by J. Marvin Weller, 176
Cited on the Marble Falls foraminifera (Chalmer L. Cooper), 263
Polycrystalline aggregates (N. Allen Riley), 460
Pontides, the (S. W. Tromp), 372

Postglacial forest succession, climate, and chronology in the Pacific northwest. By Henry P. Hansen. Review by Ernest Antevs, 450

Potentials, Electrode, and free-energy changes in geology (Carleton A. Chapman and George K. Schweitzer), 43

Pre-Cambrian unconformity, the (F. G. Snyder), 147

Primary lineation in fluvial sandstones, a criterion of current direction (W. Lee Stokes), 52

Problem of the Lipalian interval, The (F. G. Snyder), 146

Psaronius sinensis from the Omeishan basalt series in southwestern China, On the structure of (H. C. Sze), 160

Quarternary of the Texas high plains. By Glen L. Evans and Grayson E. Meade. Review by Everett C. Olson, 177

Quartz and zircon optic-axes fabric (N. Allen Riley), 472

Quartzite, Baraboo, Structural petrology of the (N. Allen Riley), 453

Raleigh graphite, The origin and importance of the (John W. Harrington), 516
description of the, 516
occurrence in beds, 518
origin of the, 519

Rankin, G. A., and H. E. Merwin. Cited on β-Al_2O_3 (Julian R. Goldsmith), 384

Raymond, P. E. Cited on the fauna of the Lower Cambrian (F. G. Snyder), 151

Read, Charles B. *Pennsylvanian floral zones and floral provinces*, 271
and Gordon H. Wood. *Distribution and correlation of Pennsylvanian rocks in late Paleozoic sedimentary basins of northern New Mexico*, 220

Reeside, J. B., and J. Gilluly. Cited on Triassic red beds (Edward L. Holt), 512

Refraction, diagrams (Walter H. Munk and Melvin A. Traylor), 2, 10, 12
effect of, on wave height, 3
of ocean waves: a process linking underwater topography to beach erosion, 1
type examples, 5

Reger, David B. Cited on Pennsylvanian cycles (Harold R. Wanless), 237
Cited on the Pennsylvanian of West Virginia (Aureal T. Cross), 288

Regional stratigraphic studies, stage of advancement of (Harold R. Wanless), 185
variations in Pennsylvanian lithology, 237

Resser, C. E. Cited on the base of the Cambrian (F. G. Snyder), 149, 150

Riedel, W. Cited on experimental shearing (N. Allen Riley), 469

Riley, N. Allen. *Structural petrology of the Baraboo quartzite*, 453

Ring-structure, The Vredefort, of South Africa (Reginald A. Daly), 125

Roberts, T. G., and R. C. Spivey. Cited on Pennsylvanian terminology in Texas (M. G. Cheney), 206

Rockbridge County, Virginia, A tectonic intrusion of shale in (Robert O. Bloomer), 48

Rocky Mountain region, floral modifications in the (Charles B. Read), 276
trough (A. J. Eardley), 312

Rocky Mountains, northern, and east central Idaho, Drainage diversion in the (Alfred L. Anderson), 61
Types of Pennsylvanian cycles in the (Harold R. Wanless), 245

Rodgers, John, P. B. King, H. W. Ferguson, and L. C. Craig. Cited on the base of the Cambrian (F. G. Snyder), 149

Rogers, W. B. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94

Rominger, J. F., and E. C. Dapples. Cited on current direction (W. Lee Stokes), 52

Ross, C. P. Cited on origin of Warm Spring Creek Basin and Lost River Range (Alfred L. Anderson), 68

Rounding of beach boulders, The rate of (J. A. Bartrum), 514

Rove slate (Upper Huronian) and associated intrusives (Karl ver Steeg), 357

Russell, H. N. Cited on meteorite craters (Reginald A. Daly), 138

Russell, W. G. Cited on the Cimarron River (Thad G. McLaughlin), 82

Saganaga granite (Laurentian) (Karl ver Steeg), 356

Sahni, B. Cited on continental drift (H. C. Sze), 165, 166

St. John, Orestes. Cited on the Cimarron River (Thad G. McLaughlin), 81

Salomon-Calvi, W. Cited on the Izmit-Erzurum fault zone (S. W. Tromp), 369

Sandstones, fluvial, Primary lineation in, a criterion of current direction (W. Lee Stokes), 52

Sawin, H. J. Amphibians from the Dockum Triassic of Howard County, Texas. Review by Everett C. Olson, 177

Scharpe, C. F. S., C. W. Thornthwaite, and E. F. Dosch. Cited on accelerated channel erosion (Thad G. McLaughlin), 89

Schenck, H. G., and S. W. Muller. Cited on lithologic boundaries (Harry E. Wheeler), 153

Schofield, S. J. Cited on basal Triassic conglomerate (A. J. Eardley), 333
Cited on Cascadis, 335

Schopf, James M. Cited on the morphology and terminology of spores (Aureal T. Cross), 290
Cited on the use of megaspores in correlation (R. M. Kosanke), 280

Schuchert, Charles. Cited on the base of the Cambrian (F. G. Snyder), 149
Cited on Cascadis (A. J. Eardley), 334
Cited on the Cordilleran geosyncline, 309
and C. O. Dunbar. Cited on *Calamites* and *Sigillaria* at Joggins, Nova Scotia (Edward L. Holt), 511
and C. O. Dunbar. Cited on the Lipalian interval (F. G. Snyder), 150

Schweitzer, George K., and Carleton A. Chapman. *Electrode potentials and free-energy changes in geology*, 43

Sea water, fluorine in (Tom. F. W. Barth), 421

Secondary-phase boundaries for the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Robert G. Smalley), 33

Sedimentary basins, late Paleozoic, of northern New Mexico, Distribution and correlation of Pennsylvanian rocks in (Charles B. Read and Gordon H. Wood), 220

Shaffer, Paul R. *Correlation of the erosion surfaces of the southern Appalachians*, 343

Shale, A tectonic intrusion of, in Rockbridge County, Virginia (Robert O. Bloomer), 48

Shand, S. J. Cited on pseudo-tachylyte (Reginald A. Daly), 131

Shenon, P. J. Cited on origin of valley depressions (Alfred L. Anderson), 68

Shepherd, E. S. Cited on the significance of fluorine in geochemistry (Tom. F. W. Barth), 420

Silicate phase of stony meteorites, The composition of meteoritic matter. I. The composition of the (Harrison Brown and Claire Patterson), 405

Small-spore genera in Illinois coal beds, distribution of (R. M. Kosanke), 280

Smalley, Robert G. *The system $NaAlSiO_4$–$Ca_2Al_2SiO_7$*, 27
Cited on the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Julian R. Goldsmith), 383, 386, 399, 400

Smith, H. T. U. Cited on terraces (Thad G. McLaughlin), 78

Snyder, F. G. *The problem of the Lipalian interval*, 146

Soda-lime-alumina system, petrologic significance of (Robert G. Smalley), 35

Sorotchinsky, C. Cited on origin of Salle du Dôme in the Grotte de Han of Belgium (E. L. Krinitzsky), 119

Spencer, L. J. Cited on meteorite craters (Reginald A. Daly), 138

Spheroidal forms, in rock exfoliation (Theodore A. Dodge), 39

Spheroids, some factors influencing the formation of (Theodore A. Dodge), 41

Spivey, R. C., and T. G. Roberts. Cited on Pennsylvanian terminology in Texas (M. G. Cheney), 206

Spore floras of the Pennsylvanian of West Virginia and Kentucky (Aureal T. Cross), 285

Spores, descriptive terminology of, 291

Stark, J. T. Cited on rhyolite in the Baraboo region (O. E. Gram), 427

Stearns, H. T., and G. A. Macdonald. Geology and ground water resources of the island of Hawaii. Review by Rollin T. Chamberlin, 378

Steers, J. A. The coastline of England and Wales. Review by Leland Horberg, 378

Stenzel, H. B. Decapod crustaceans from the Cretaceous of Texas. Review by J. Marvin Weller, 177

Stokes, W. Lee. *Primary lineation in fluvial sandstones: a criterion of current direction*, 52

Stose, G. W., and A. I. Jonas. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94, 102, 106

Stout, W. Cited on Pennsylvanian cycles (Harold R. Wanless), 237

Strahler, A. N., and D. L. Babenroth. Cited on faulting (Sheldon Judson), 172

Stratigraphic studies, regional, stage of advancement of (Harold R. Wanless), 185

Stratigraphy, and the erosion surfaces of the southern Appalachians (Paul R. Shaffer), 351
of the Lower Pennsylvanian coral-bearing strata of Texas. By F. B. Plummer. Review by J. Marvin Weller, 176

Stream flow (Thad G. McLaughlin), 79

Streams, tributary, accelerated erosion in, 86

Structural features, of the rocks north of Lake Superior (Karl ver Steeg), 356

Structural petrology of the Baraboo quartzite (N. Allen Riley), 453

Structural petrology of the Pine Creek area, Dickinson County, Michigan (James W. Higgins), 476

Structural unconformity, primary, of the Pine Creek area, 476

Structural units, main, of the Anatolian orogenic belt, A tentative classification of the (S. W. Tromp), 362

Structure, geologic, The influence of, on the drainage pattern in northeastern Minnesota (Karl ver Steeg), 353
of Psaronius sinensis from the Omeishan basalt series in southwestern China, On the (H. C. Sze), 160

Structures, Large-scale superficial—a discussion (Sheldon Judson), 168

Sturgeon, M. T. Cited on forms of *Mesolobus* (M. G. Cheney), 217

Submarine canyons (Walter H. Munk and Melvin A. Traylor), 6

Submarine relief off the Norwegian coast with bathymetrical map in seven sheets of the Norwegian coastal waters and adjoining areas, The. By Olaf Holtedahl. Review by Leland Horberg, 58

Submarine ridges (Walter H. Munk and Melvin A Traylor), 7

Summary of classification of the Pennsylvanian formations of Texas, with special reference to the Lower Pennsylvanian of the Llano region (Frederick B. Plummer), 193

Swinnerton, A. C. Cited on development of caverns (E. L. Krinitzsky), 116

Symposium on Pennsylvanian problems, Introduction to (Harold R. Wanless), 184

System $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$, the binary (Julian R. Goldsmith), 387
$CaAl_2Si_2O_8$–$NaAlSiO_4$, the, 384
$CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$, The, 381
$CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$, liquidus relations, 388
nonternary nature of, 388

System—*continued*
petrologic significance of, 403
$NaAlSiO_4$–$Ca_2Al_2SiO_7$, The (Robert G. Smalley), 27
$NaAlSiO_4$–$Ca_2Al_2SiO_7$, the (Julian R. Goldsmith), 386
thermal data, 387
Sze, H. C. *On the structure of Psaronius sinensis from the Omeishan basalt series in southwestern China*, 160

Taylor, J. H., and S. E. Hollingworth. Cited on cambers (Sheldon Judson), 170
Tectonic features of the Utley meta-rhyolite (O. E. Gram), 427
Tectonic forms of China, On major. By T. K. Huang. Review by Rollin T. Chamberlin, 59
Tectonic Intrusion of shale in Rockbridge County, Virginia, A (Robert O. Bloomer), 48
Tennessee, history of zinc mining in (Arthur T. Allen), 412
The Longview member of the Kingsport formation, 412
Tertiary volcanics, distribution pattern of (Alfred L. Anderson), 71
Texas, north-central, Pennsylvanian classification and correlation problems in (M. G. Cheney), 202
Summary of classification of, with special reference to the Lower Pennsylvanian of the Llano region (Frederick B. Plummer), 193
Thermal data for the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (Robert G. Smalley), 30
Thermal-equilibirum relationships. *The system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$* (Julian R. Goldsmith), 381
Thompson, M. L. Cited on fossils in the Putnam Hill and Vanport limestones (M. G. Cheney), 217
Cited on the *Millerella* zone (Chalmer L. Cooper), 262
Thornthwaite, C. W., C. F. S. Scharpe, and E. F. Dosch. Cited on accelerated channel erosion (Thad G. McLaughlin), 89
Tilghman, B. C. Cited on overtilting of rim of Canyon Diablo crater (Reginald A. Daly), 140
Tilley, C. E. Cited on carnegieite solid solutions (Julian R. Goldsmith), 400
Cited on sodium metasilicate (Robert G. Smalley), 34
Time indices, unconformities as, a false concept (Harry E. Wheeler), 154
Topography, of the area underlain by the Rove slate and associated intrusives (Karl ver Steeg), 358
underwater, to beach erosion, Refraction of ocean waves: a process linking (Walter H. Munk and Melvin A. Traylor), 1
Toros orogen, the (S. W. Tromp), 372
Traylor, Melvin A.and Walter H. Munk. *Refraction of ocean waves: a process linking underwater topography to beach erosion*, 1
Triassic, Upper, of western Colorado, Upright trunks of Neocalamites from the (Edward L. Holt), 511
Tributary streams, accelerated erosion in (Thad G. McLaughlin), 86
Triletes, genus, section Aphanozonati, discussion of the (Aureal T. Cross), 295
section Auriculati, discussion of the, 303
section Lagenicula, discussion of the, 304
section Triangulati, discussion of the, 299
Tromp, S. W. *A tentative classification of the main structural units of the Anatolian orogenic belt*, 262
Turkey. *A tentative classification of the main structural units of the Anatolian orogenic belt* (S. W. Tromp), 362
Tuttle, O. F. Cited on planes of inclusions (N. Allen Riley), 472

Umpleby, J. B. Cited on dissected uplands (Alfred L. Anderson), 63
Unconformities as time indices, a false concept (Harry E. Wheeler), 154
Underwater topography to beach erosion, Refraction of ocean waves: a process linking (Walter H. Munk and Melvin A. Traylor), 1
Upright trunks of Neocalamites from the Upper Triassic of western Colorado (Edward L. Holt), 511
Utley meta-rhyolite, Tectonic features of the (O. E. Gram), 427
chilled zones in the, 435
extrusive character of, 431
flow structure in the, 434
gash veins in the, 435
general geology of the, 429
inclusions in the, 434
joint system in the, 436
layers of spheroids in the, 431
structural geology of the, 431
structure of, interpretation of the, 436
volcanic breccia in the, 433

ver Steeg, Karl. *The influence of geologic structure on the drainage pattern in northeastern Minnesota*, 353
Virginia. *A fault-plane cavern* (E. L. Krinitzsky), 107
central, The Catoctin formation in (Robert O. Bloomer and Richard R. Bloomer), 94
A tectonic intrusion of shale in Rockbridge County (Robert O. Bloomer), 48
Volcanic archipelago (A. J. Eardley), 334
Vredefort ring-structure of South Africa, The (Reginald A. Daly), 125
description of, 126
five theories of origin of, 132
as a meteorite scar, 137

Walcott, Charles D. Cited on the origin of the term "Lipalian" (F. G. Snyder), 146
Cited on sections in British Columbia (A. J. Eardley), 313
Wanless, Harold R. *Introduction to Symposium on Pennsylvanian problems*, 184
Regional variations in Pennsylvanian lithology, 237
Pennsylvanian geology of a part of the southern Appalachian coal field. Review by J. Marvin Weller, 181

Cited on Kanawha black flint and Roaring Creek sandstone in West Virginia (M. G. Cheney), 216
Washington, H. S., and F. W. Clarke. Cited on the amount of fluorine in the earth's crust (Tom. F. W. Barth), 420
Wasserstein, B. Cited on fluorine concentration in granite (Tom. F. W. Barth), 421
Watson, F. G. Cited on meteorite scars (Reginald A. Daly), 138
Wave refraction and beach erosion (Walter H. Munk and Melvin A. Traylor), 22
bottom topography and, 20
meteorology and, 19
physical nature of, 2
theory of, 24
Wave trains, crossing of, north of Scripps Canyon, 19
Waves, ocean, Refraction of: a process linking underwater topography to beach erosion (Walter H. Munk and Melvin A. Traylor), 1
Weathering, chemical, An example of exfoliation caused by (Theodore A. Dodge), 38
Weaver, C. E. Cited on the Stevens series of Washington (A. J. Eardley), 323
Wedd, C. B. Cited on dip-and-fault structure (Sheldon Judson), 172
Weidman, Samuel. Cited on petrography of pre-Cambrian rocks in Wisconsin (O. E. Gram), 430
Cited on the structure of the Baraboo quartzite (N. Allen Riley), 455
Weller, J. Marvin. *Invertebrates in Pennsylvanian correlations*, 254
Review of: Charles Darwin and the voyage of the Beagle. Edited by Nora Barlow, 452
Review of: Decapod crustaceans from the Cretaceous of Texas. By H. B. Stenzel, 177
Review of: Description of Lower Pennsylvanian corals from Texas and adjacent states. By Raymond C. Moore and Russell M. Jeffords, 176
Review of: Pennsylvanian geology of a part of the southern Appalachian coal field. By Harold R. Wanless, 181
Review of: Report of the committee on marine ecology as related to paleontology. By Harry S. Ladd, chairman, 180
Review of: Smaller foraminifera in the Marble Falls, Smithwick, and Lower Strawn strata around the Llano uplift in Texas. By Helen Jeanne Plummer, 176
Review of: Stratigraphy of the Lower Pennsylvanian coral-bearing strata of Texas. By F. B. Plummer, 176
Review of: The Wilberns Upper Cambrian graptolites from Mason, Texas. By Charles E. Decker, 176
Cited on rhythmic pattern of Pennsylvanian beds (Harold R. Wanless), 237
C. O. Dunbar, L. G. Henbest, and. Cited on fusulinid correlation (M. G. Cheney), 217
and Everett C. Olson. Review of: Contributions to geology, 1944, 176
West Virginia and Kentucky, Spore floras of the Pennsylvanian of (Aureal T. Cross), 285
Westgate, L. G., and A. Knopf. Cited on thickness of sediments (A. J. Eardley), 314
Wheeler, Harry E. *Base of the Cambrian system*, 153
Cited on the Cambrian of Nevada (F. G. Snyder), 148
White, C. H. Cited on primeval sea (Tom. F. W. Barth), 421
Wilberns Upper Cambrian graptolites from Mason, Texas, The. By Charles E. Decker. Review by J. Marvin Weller, 176
Williams, G. H. Cited on the Catoctin formation (Robert O. Bloomer and Richard R. Bloomer), 94
Williams, H. S. Cited on correlation of faunas (Charles B. Read), 271
Williams, J. S. Cited on source of sandstones and red beds (A. J. Eardley), 314
Wilmarth, M. Grace. Cited on original definition of the Cambrian system (Harry E. Wheeler), 153
Wilson, R. R. Cited on the Cimarron River (Thad G. McLaughlin), 84
Winchell, A. N. Cited on melilite molecules (Robert G. Smalley), 35
Cited on melilites (Julian R. Goldsmith), 401
Wind gaps, in the northern Rocky Mountains (Alfred L. Anderson), 70
Wisconsin. *Structural petrology of the Baraboo quartzite* (N. Allen Riley), 453
Tectonic features of the Utley meta-rhyolite (O. E. Gram), 427
Wislizenus, A. Cited on Cimarron River (Thad G. McLaughlin), 81
Wood, Gordon H., and Charles B. Read. *Distribution and correlation of Pennsylvanian rocks in late Paleozoic sedimentary basins of northern New Mexico*, 220
Wright, F. E. and C. W. Cited on structural trends of the Triassic conglomerate (A. J. Eardley), 334
Wright, H. E. Cited on faulting (Sheldon Judson), 172

Zaruba-Pfeffermann, Quido. Cited on Toreva-block sliding (Sheldon Judson), 172
Zeuner, Frederick E. Dating the past: an introduction to geochronology. Review by Ernst Antevs, 527
The Pleistocene period: its climate, chronology and faunal successions. Review by Ernst Antevs, 446
Zinc mining in Tennessee, history of (Arthur T. Allen), 412
Zircon, quartz and, optic-axes fabric (N. Allen Riley), 472
Zuidema, Henry P. *Discovery of letters by Lyell and Darwin*, 439

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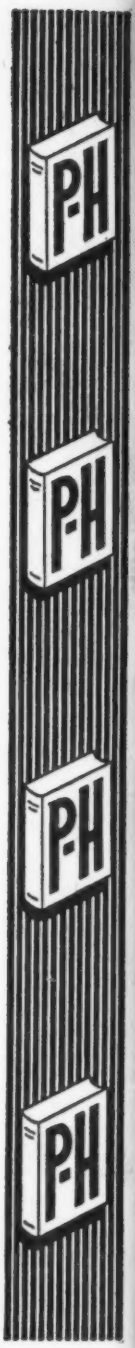
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